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# Economic Geology

and the

## Bulletin of the Society of Economic Geologists

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# REMANENT MAGNETISM AND THE ORIGIN OF HARD HEMATITES IN PRECAMBRIAN BANDED IRON FORMATION

W. H. GROSS AND D. W. STRANGWAY

ABSTRACT

Concentrations of hard hematite in iron formation might be formed by syngenetic, metamorphic, magmatic or surface leaching processes. It is usually of practical value to know the origin of hematite ores but the origin is often obscure and difficult to determine by classical geological methods.

Hematite has relatively stable magnetic properties so it was thought that an analysis of the directions of remanent magnetism might be useful in determining the origin of hematite ore bodies. To test this possibility, the remanent magnetic directions were determined from samples of two ore bodies that occur in a highly deformed Precambrian iron formation at Fort Gouraud, Mauritania. The results suggest that one ore body was formed in Precambrian time, when the iron formation was flat lying, while the second ore body was formed by surface leaching with at least some of the solution and redeposition of the hematite occurring in recent time. The origin of these ore bodies as interpreted from the remanent magnetic data is consistent with the geological field relationships.

## INTRODUCTION

AT present the most important sources of iron ore are the hematite concentrations that occur in banded iron formations of Precambrian age. Iron formations are iron-rich, chemical sediments in which the iron minerals (mainly hematite) are interbanded with any combination of silica, carbonate, or silicate minerals. Normally these sediments contain from 15 to 30 percent iron and are not mined as direct shipping ore. In some places natural enrichments of hematite occur within the iron formation and raise the iron content from 45 percent up to about 70 percent; it is these concentrations that make the important bodies of direct shipping ore.

As pointed out recently by Park (9) there is considerable confusion about the origin of hard hematite ore bodies in iron formation. Although it is realized by geologists that hematite could form within the iron formation in a number of quite different ways there seems to be little agreement on the importance of each ore-forming process or on criteria for recognizing one ore-forming process from another.

It is possible to think of at least four separate natural processes that could concentrate hematite in iron formation. These processes might be summarized as follows:

(1) *Syngenetic Hematite Ore.*—Normally iron formation is composed of alternating bands of iron-rich and gangue-rich material. If during sedimentation there is an interruption in the deposition of gangue minerals then any continued deposition of the iron-rich constituents may be sufficient to form a syngenetic concentration of iron within the formation. Various other processes might be visualized as forming syngenetic ore and some of these are shown on Diagram A.

(2a) *Metamorphic Hematite Ore.*—Iron formation (approximately 30% iron) after deposition might be subjected to one or more periods of tectonic activity. During these disturbances the beds are usually folded and faulted and the minerals metamorphosed. Under conditions of higher temperatures and pressure and in the presence of connate or ground water, the components of the iron formation may be mobilized and differential flow of the constituents could separate the hematite from the gangue minerals. The hematite, for example, could either migrate to favorable structural environments for deposition or the gangue minerals could migrate, leaving hematite ore as a residual concentration (Diag. A).

(2b) *Magmatic Hematite Ores.*—Igneous intrusions near the iron formation could either supply additional iron to the formation to form ore, or solutions emanating from a magma could leach the gangue from the iron formation leaving residual concentrations of hematite.

(3) *Lateritic Hematite Ores.*—Iron formation exposed at the surface can be leached of gangue minerals during weathering and the residual hematite could accumulate on the surface in sufficient quantities to form ore bodies. Generally the hematite formed by this process is soft and earthy but may be capped with a crust of hard hematite material.

Undoubtedly these simple concepts of the origin of hematite ores in iron formation may be modified and combined in several ways. For example, soft shallow laterites may be metamorphosed and folded so that they become deep hard ores. It is probably true, however, that if the origin could be relegated to one or a combination of the four processes outlined above the knowledge would be of value not only in developing known ore bodies but also in locating new deposits in the same district. If the hematite ore could be shown to be syngenetic then the ore bodies could be expected to be concordant with the banding in the iron formation and to be relatively continuous along strike. These ore bodies would be independent of such secondary structures as folding and faulting and the broad approach to the exploration for new ore would be to try to reconstruct the original depositional basin and to think of ore formation in terms of the vicissitudes of primary sedimentation within this basin.

If the hematite could be shown to have been concentrated by metamorphic or magmatic processes then the developer should look for a structural control for the accumulation of the hematite. Secondary structures such as folds

and faults are important in these ore-forming processes for they may act not only as channelways through which solutions can move but also as structural traps which can house the ore. Knowledge of the structural environment is, therefore, a useful guide in the exploration for new ore bodies of this type.

Finally, if the ore could be shown to have been concentrated by weathering and leaching, then the explorer should realize that the depth and the shape of the ore bodies will be related to the topography that existed at the time of ore formation.

There seems to be little doubt that knowledge of their origin would be useful in the exploration and development of hematite ore bodies. The chief problem, particularly in the early stages of development, is in recognizing which ore type is being explored. There are, of course, a great number of geological and laboratory criteria that are helpful. The present paper attempts to show that a study of the remanent magnetism of hematite, which is a mineral with a high degree of magnetic stability, might be useful in helping to recognize the origin of hematite ore bodies.

## REMANENT MAGNETISM

*General.*—Rocks containing iron oxides have two main types of magnetization—induced and permanent (or remanent). Many rocks have a remanent magnetization that is extremely stable and may survive through long periods of geological time without change in direction. This magnetization in many cases appears to reproduce the direction of the earth's field at the time of formation of the rock (1, 8).

In recent years the remanent magnetism of rocks has been widely studied, and the results from many continents are now becoming available. It appears that the earth's magnetic field has wandered considerably during geological time. Therefore, the purpose of many of the studies is to use the direction of permanent magnetism, as found in the rocks of various ages, to reconstruct the position of the earth's magnetic pole through geologic history (6, 10). If the history of the polar wandering could be unravelled it would be helpful in giving rocks a relative and perhaps even an absolute age.

Remanent magnetism has also been used in structure studies, particularly in folded sediments. Furthermore, although very little work has been done in this direction, remanent magnetism may be an important tool in helping to reveal the origin of ores containing iron minerals.

*Types of Remanent Magnetism.*—There are several ways in which rocks may acquire a permanent magnetic moment. Most igneous rocks, for example, acquire a permanent magnetism when cooled from above the Curie temperature of the magnetic minerals present. This magnetization can be very large and may be stable. The Curie temperature of magnetite is 585° C, and that of hematite is about 670° C. Thus hematite ore bodies that were formed above 670° C or that were subsequently heated above this temperature could be expected to have a permanent magnetic moment that would agree with the direction of the magnetic field that existed at the time of last heating.

If it could be shown that a rock was formed below the Curie temperature of its magnetic minerals and was never heated above this temperature by later geological processes, then any remanent magnetism found to be present

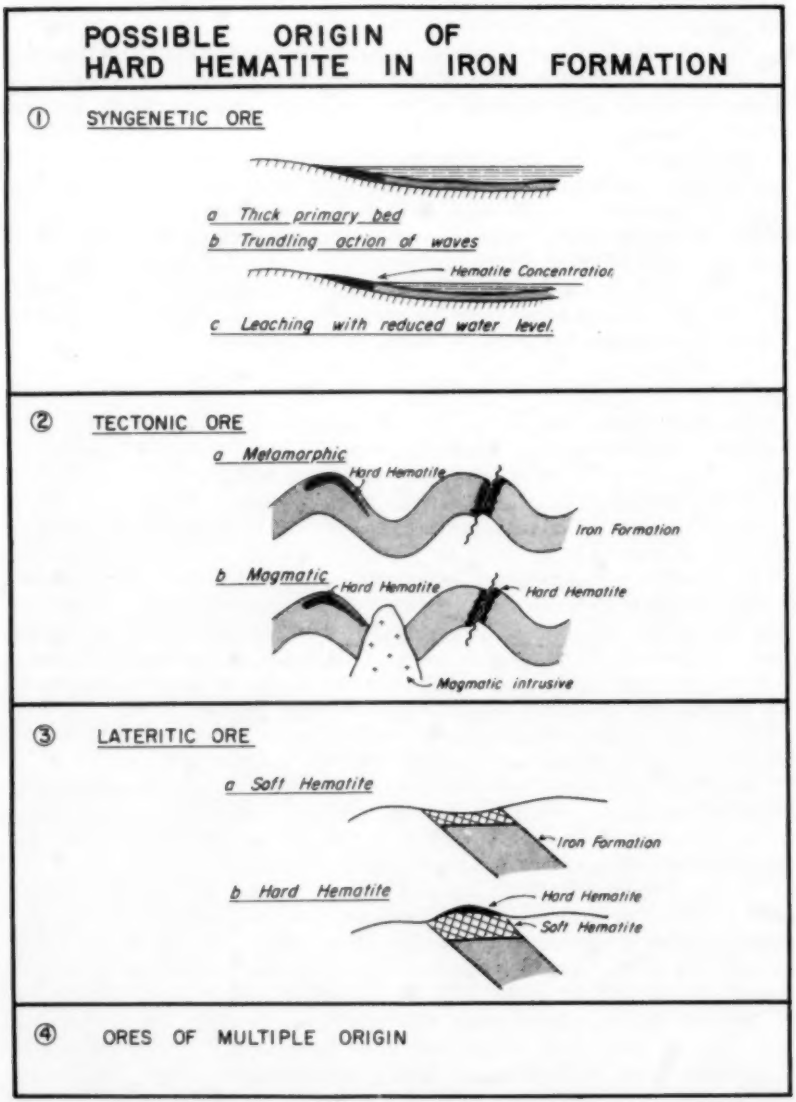



DIAGRAM A.

must have formed in some other way. In sedimentary environments a permanent magnetization may be acquired by the rock as the grains of magnetic minerals settle through the water and align themselves parallel to the earth's field. King (*7*) has made a study of this process in the laboratory and he has shown that there is a tendency for the acquired magnetization to be somewhat flatter than the applied field. However, under favorable conditions sedimentary formations commonly show fairly uniform directions of magnetization that appear to be closely related to the direction of the earth's field at the time of deposition.

Unfortunately, there appear to be a number of other ways by which the remanent magnetic direction of the rocks may be changed or even obliterated. Spurious results may be caused by an increase in pressure and temperature during metamorphism, by chemical alteration of the magnetic mineral, or even by the effects of lightning on surface exposures.

*Tests of Magnetic Stability.*—Naturally it is important to know whether the remanent magnetism being measured in a rock is either primary, formed at the same time as the rock; or secondary, formed at some time later than the rock. Unfortunately, there is no truly satisfactory method of identifying primary magnetism because it is impossible to devise adequate tests for long term stability in the short time interval involved in any laboratory experiment. The most satisfactory test, as suggested by Graham (3), is the self-consistency of data collected from the field where various geological units show a uniform magnetization, which is different in direction from that of the adjacent geologic units and different from that of the present earth's field. Another good test of stability is the measurement of directions of magnetization on the two limbs of a fold. If the same lithologic unit has two different directions of magnetization on its two limbs and if these two directions coincide when the bed is "unfolded," the remanent magnetism must be stable and must have been acquired before folding. If the original position of the bed is known and if the direction of magnetization corresponds with that of the earth's field at the time of the original sedimentation, then the remanent magnetization was undoubtedly acquired during sedimentation.

*Remanent Magnetism in Iron Formation.*—The cause of deposition and the composition of the original iron minerals in an iron formation are uncertain. If the principal iron mineral is hematite, it could be visualized as settling out as fine-grained material in relatively still water. Under these conditions it is probable that the tiny hematite grains align themselves with the earth's magnetic field at the time of deposition.

Hematite is known to be magnetically stable, particularly when it is fine grained. For example, the coercive force, which is the applied field required to remove saturation magnetization, is about 760 oersteds. This does not mean that the original direction of magnetization acquired by the hematite could not change if the rocks were subsequently heated above the Curie temperature or if the rocks were to undergo extensive dynamic metamorphism or alteration. But it does mean that hematite is far more likely to preserve its direction of magnetization than other natural magnetic min-

erals. For this reason studies involving hematite as the principal magnetic mineral often show more self-consistency than other studies.

It seems clear that the primary direction of magnetization in the hematite of an iron formation should be retained during folding provided the temperature of the rocks does not rise above 670° C. If, after folding, some gangue minerals in the iron formation were replaced by secondary hematite, then the direction of magnetization acquired by this new material would normally be independent of the magnetic direction already present in the primary hematite bands. The explanation is that an infinite magnetic sheet as mani-

FIG. 1.

fested by the primary bands in the iron formation does not distort the applied field outside that sheet (11). This means that the primary hematite formed during sedimentation when the beds were flat would have one direction of magnetization while any hematite introduced or redeposited within the iron formation subsequent to folding, metamorphism or igneous activity, would have another. The remanent magnetic directions in hematite formed by surface weathering would be expected to have a random orientation due to slumping. If a hard hematite crust is formed by recrystallization of some of the surface hematite there may be a preferred orientation in the direction of the earth's field at the time of this consolidation and recrystallization.

## THE FORT GOURAUD HARD HEMATITE ORE BODIES

The Fort Gouraud hard hematite ore bodies are located in French West Africa as shown in Figure 1. The hematite occurs entirely within a Precambrian banded iron formation. The distribution of the iron formation and the location of the principal ore bodies within it are shown in Figure 2.

*General.*—The F'Derik body is composed almost entirely of hard hematite which grades upwards of 65 percent total iron and which is housed in a Precambrian banded iron formation. The hematite is of two types; one is essentially massive, is concordant with the bedding, and is known to extend to a depth of at least 750 feet below the present surface with no detectable

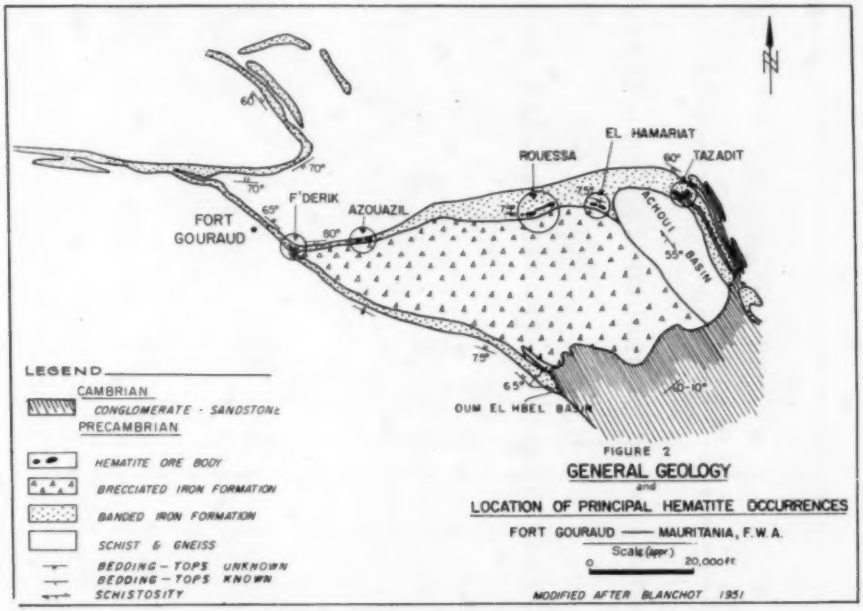



FIGURE 2
GENERAL GEOLOGY
and
LOCATION OF PRINCIPAL HEMATITE OCCURRENCES
FORT GOURAUD — MAURITANIA, F.W.A.
Scale (appr.) 0 20,000 ft.

## THE F'DERIK ORE BODY

change in its physical or chemical properties (2). The distribution of this ore is shown in Figure 3 and is marked "deep hematite ore." The other type of hematite, although similar in appearance to the deep hematite, is not concordant with the bedding and does not extend to depth. It is marked in Figure 3 as "surface hematite ore."

It may be seen in Figure 3 that the deep ore and the surrounding iron formation have been folded into an elongated S-shaped structure that plunges north and which, in the horizontal plane, has an angle of about 37 degrees

TABLE I
MAGNETIC MEASUREMENTS ON F'DERIK ORE BODY

| Sample | Declination | Inclination | Avg. Rem. Mag. x $10^3$ oe | Avg. Susc. x $10^3$ cgs/cc | Q = $\frac{J}{xH}$ * |
|---|---|---|---|---|---|
| 2 | 26 | 60 | .145 | .330 | .9 |
| 3 - 2 | 286 | 58 | - | | |
| 4 - 1 | 59 | 6 | 1.390 | 1.340 | 2.1 |
| - 2 | 49 | 11 | | | |
| 5 | 282 | 14 | 3.02 | .700 | 8.5 |
| 6 | - | 59 | .490 | .397 | 2.5 |
| 8 | 48 | - 8 | 1.22 | .890 | 2.8 |
| 9 | 47 | -29 | | | |
| 10 | 12 | -12 | .655 | 1.720 | 0.8 |
| 11 | 183 | - 8 | 5.43 | 1.140 | 9.6 |
| 12 | 314 | 54 | 1.10 | .420 | 5.2 |
| 20 - 1 | 13 | 12 | 4.23 | 1.000 | 8.5 |
| - 2 | 13 | 12 | | | |
| - 3 | 90 | -14 | | | |
| - 4 | 209 | - 1 | | | |
| - 5 | 183 | -22 | | | |
| - 6 | 9 | 16 | | | |
| - 7 | 17 | -10 | | | |
| - 8 | 27 | - 1 | | | |
| - 9 | 26 | 1 | | | |
| 22 - 1 | 238 | 26 | 1.38 | .580 | 4.8 |
| - 2 | 180 | 12 | | | |
| 23 | 38 | 22 | .995 | .580 | 3.3 |
| 24 - 1 | 43 | 75 | 1.10 | .480 | 4.6 |
| - 2 | 33 | 36 | | | |
| - 3 | 37 | 30 | | | |
| - 4 | 14 | 52 | | | |
| - 5 | 21 | 46 | | | |
| - 6 | 32 | 34 | | | |
| 25 | 68 | - 2 | 17.0 | .310 | 109. |
| 26 - 1 | 97 | -36 | 8.57 | .400 | 43. |
| - 2 | 100 | -42 | | | |
| - 3 | 103 | -64 | | | |
| - 4 | 112 | -46 | | | |
| 27 - 1 | 33 | 28 | .617 | .510 | 2.4 |
| - 2 | 33 | 28 | | | |
| - 3 | 36 | 28 | | | |
| - 4 | 36 | 27 | | | |
| - 5 | 33 | 26 | | | |
| - 6 | 37 | 26 | | | |
| - 7 | 34 | 22 | | | |
| - 8 | 36 | 28 | | | |
| - 9 | 35 | 28 | | | |
| -10 | 38 | 23 | | | |
| -11 | 35 | 21 | | | |
| -12 | 34 | 26 | | | |
| 28 - 1 | 284 | 25 | 62.5 | 1.650 | 76. |
| - 2 | 287 | 32 | | | |
| - 3 | 295 | 25 | | | |
| - 4 | 280 | 20 | | | |
| 29 - 1 | 32 | 25 | 4.72 | .950 | 10 |
| - 2 | 21 | 43 | | | |
| - 3 | 34 | 11 | | | |
| - 4 | 42 | 26 | | | |
| - 5 | 42 | 21 | | | |
| - 6 | 70 | - 4 | | | |

*Q = $\frac{J}{xH}$ is the ratio of remanent to induced magnetization where x is the susceptibility and H is the applied field.

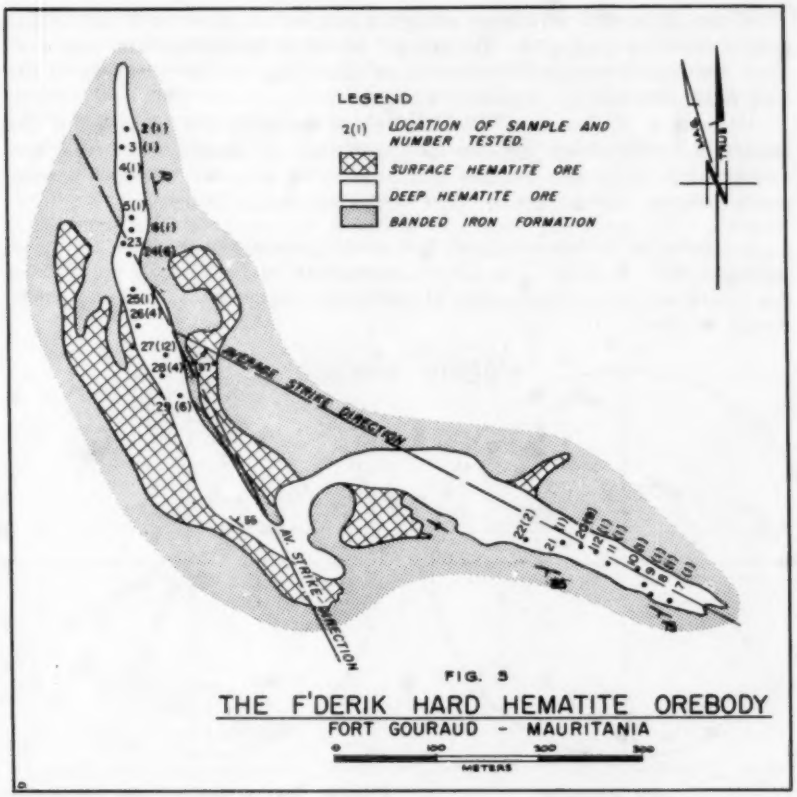



FIG. 3

THE F'DERIK HARD HEMATITE OREBODY

FORT GOURAUD - MAURITANIA

between the two limbs. The south limb dips nearly vertical and the north limb dips about 65° E.

*Collection and Measurement of Samples.*—Twenty-nine oriented samples located as shown in Figure 3 were collected from the F'Derik ore body. These samples were cut into one-inch cubes and spun in a remanent magnetometer measuring both the direction and intensity of permanent magnetism (4). The results of these measurements are given in Table 1. This table shows the average magnetic susceptibility of each sample, measured in fields of about one oersted, as well as the ratio of remanent to induced magnetization (Q).[1] It is interesting to note from the results in Table 1 that although the remanent magnetism is weak, the Q ratio is very high, indicating that the remanent magnetism is the dominant form of magnetism present in these samples.

[1] $Q = J/xH$ = the ratio of remanent to induced magnetization where x is the susceptibility and H is the applied field.

When the sample was large enough a number of cubes were cut so that results could be compared. Because of errors in orientation and measurement, the results are probably not closer than 10° and interpretation of the data must be based on statistical methods.

*Discussion of Results, F'Derik.*—Before analyzing the meaning of the results in terms of ore genesis it is necessary to determine whether the magnetism is stable and whether it is primary or secondary. The following points indicate that the greater part of the magnetism is primary.

i) Geological evidence suggests that the temperature of the hematite never exceeded 400° C (2). The Curie temperature of hematite is well above this figure so that reorganization of remanent magnetism by metamorphism should be small.

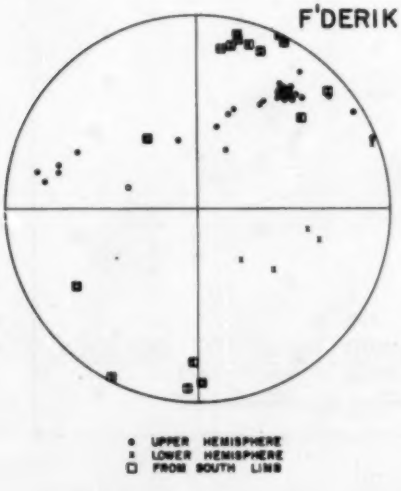



FIG. 4A. Direction of remanent magnetization, Schmidt stereographic net (north seeking poles). Vertical line points N in 4A and 4B.

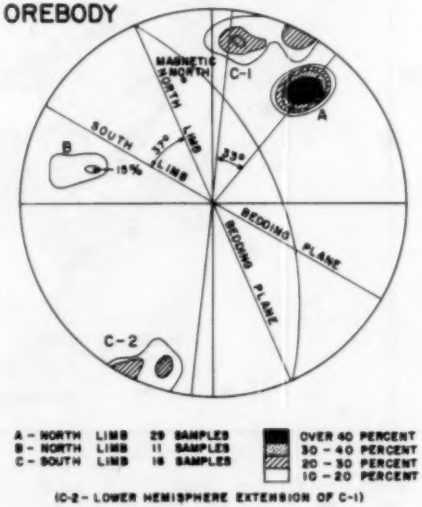



FIG. 4B. Equal area plot of north seeking magnetic poles (upper hemisphere).

ii) A few samples were subjected to A.C. demagnetizing tests in fields up to 300 oersteds. Little or no change in the magnetic direction was found indicating that the magnetization is fairly stable. This was to be expected for fine-grained hematite.

iii) A Schmidt stereographic projection and an equal area contour plot of the magnetic directions for each limb of the F'Derik fold are shown in Figure 4.[2] Since the grouping of remanent magnetic directions at an angle

[2] When the magnetic directions are plotted on a Schmidt stereographic net, several samples were found to have reversed magnetization. Such magnetic reversals occur in rocks of all

to the present earth's field is an important indicator of magnetic stability, the three concentrations of magnetic readings shown at A, B and C in Figure 4B, which make wide angles with the present magnetic north, indicate very strongly that the remanent magnetization is unaffected by the present earth's field and that the magnetization is probably stable.

iv) The equal area plot in Figure 4B also shows the strike and dip of the two limbs of the F'Derik fold. The angular difference between the two limbs is about 37°. The angle between the two principal concentrations of magnetic directions marked A and C in Figure 4 is about 33°, which is close to the angular difference between the two limbs. Now if the limbs are brought into the same vertical plane by rotating the north limb through 25° to make it vertical and then turning it through 37° to bring it into the

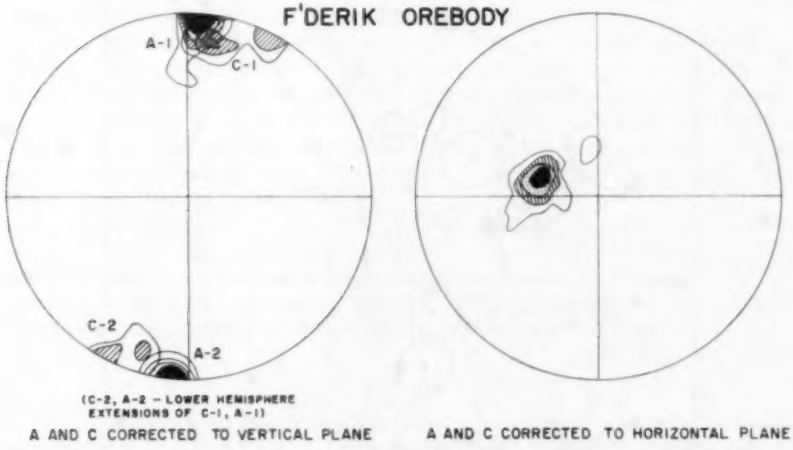



FIG. 5. Equal area contour plots corrected to vertical and horizontal planes. Vertical line is N.

same plane as the south limb, then the two sets of magnetic results, A and C, very nearly overlap, as shown in Figure 5A. If the entire ore zone is rotated to the horizontal, all the magnetic results contoured together show a good grouping of directions, as shown in Figure 5B. The coincidence of the magnetic directions from the two limbs is strong evidence that the magnetization is stable and primary, the direction of magnetism having been fixed in the hematite ore before the beds were folded.

ages and from all parts of the world and are one of the major problems in paleomagnetic studies. One explanation is that the earth's magnetic field reversed itself from time to time (10). Very often, however, results are found which cannot be explained on the basis of reversals of the earth's field and some physico-chemical process of unknown origin must be involved. In this study, for example, two cubes from a single hand specimen showed opposite polarity. Where this occurs the results have been plotted in the negative direction on the assumption that the reverse polarization was somehow established by the same field that caused the normal polarization.

The position of the magnetic poles at the time of ore formation, when the beds were flat, may now be analyzed. The declination, of course, is meaningless unless the beds quite fortuitously were unfolded to the horizontal plane in such a way as to reorient the ore bed into its original position. Unfortunately, there is no way of knowing whether or not this was done.

The inclination, on the other hand, indicates that the earth's magnetic field dipped steeply and that the magnetic pole was probably close to the Fort Gouraud area at the time of ore formation. A compilation of paleomagnetic pole positions available for Precambrian rocks in various parts of the world is shown in Figure 6 (6). The figure shows that several of the pole positions in the Precambrian are close to the present equator and are, therefore, in rough agreement with the F'Derik results.

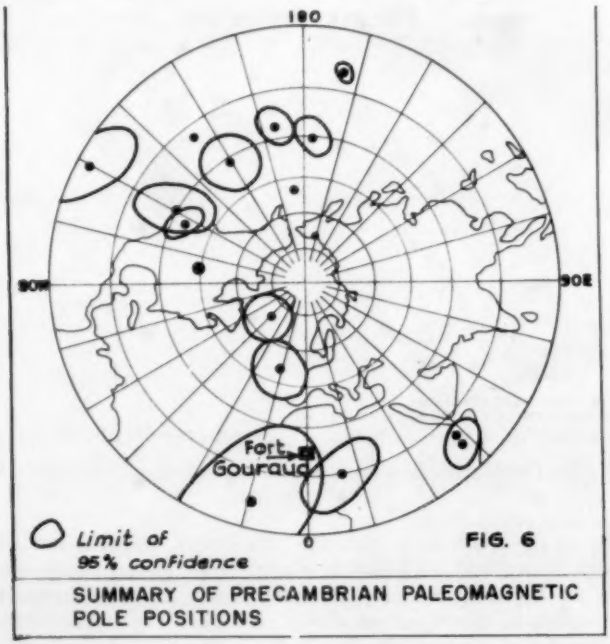



FIG. 6

SUMMARY OF PRECAMBRIAN PALEOMAGNETIC POLE POSITIONS

*Geological Interpretations—F'Derik.*—Most of the pole positions for each limb of the F'Derik fold fall on two peaks, at A and C, in Figure 4B. It has been shown by a number of independent approaches that the remanent magnetism for A and C concentrations must be stable and was probably formed before the hematite was folded. This is interpreted to mean that the bulk of the "deep hematite ore" at F'Derik is either predominantly syngenetic and represents a concentration of iron within the original basin during sedimentation or that it was formed by the leaching of the iron formation when the beds were flat lying (neglecting any primary dip).

The third and minor peak B (Fig. 4B) results principally from samples taken from the north limb and is at a considerable angle to the primary remanent magnetism A and C. It is known that the hematite in the iron formation was somewhat mobile because it is commonly found filling small fissures that cut across the banding of the formation. Furthermore, although the overall trend of the hematite lens is roughly parallel to the banding in the adjacent iron formation, the contacts of the lens may be irregular and the hematite appears, in a number of places, to have pervaded or cut gently across the iron formation.

It is believed that the "B" direction in Figure 4B represents a secondary magnetization that could have developed in the ore when some of the hematite was remobilized during folding and metamorphism or when additional but relatively minor quantities of hematite were introduced into the ore zone from some outside source.

The origin of the F'Derik ore body appears to be multiple and complex. Most of the hematite seems to have formed when the beds were relatively flat but there seems to have been either a rearrangement or addition of hematite into the ore zone at a later time. In addition, a relatively shallow capping of hematite covers the iron formation on the flanks of the principal lens, which appears to have formed after folding. No samples were collected from this zone.

It is uncertain to what extent the original hematite was enriched by later processes, but if the bulk of the ore were formed prior to folding then exploration for additional ore of the F'Derik type would depend not on the location of secondary structures, such as folds and faults, but on primary structures such as might exist around the original basin of deposition.

### THE ROUESSA WEST ORE BODY

The Rouessa West ore bodies are predominantly soft hematite that has a capping of a few feet of hard hematite (Fig. 7). Recent uplift has caused erosion to cut through the ore body down to the underlying iron formation so that the ore zone can be seen virtually in three dimensions. Figure 7 shows that there are a few lenses of hard hematite that are roughly concordant with the bedding of the underlying iron formation. The bulk of the ore is not concordant; it only extends to a shallow depth below the crest of the hills and appears to sit like a mushroom that makes a topographic high on top of the iron formation.

Five samples, from which 16 cubes were cut, were taken from the hard hematite capping. The location of these samples is shown on Figure 7 and the magnetic measurements are given in Table 2. There are insufficient data from which to draw definite conclusions but the results appear to show a random distribution of the magnetic directions with a certain preference towards the present magnetic north (Fig. 8). The random distribution might be expected from residual accumulations of hematite formed by surface leaching when the gangue minerals are removed and the hematite bands slump together in a fairly disorganized mass. Any remanent mag-

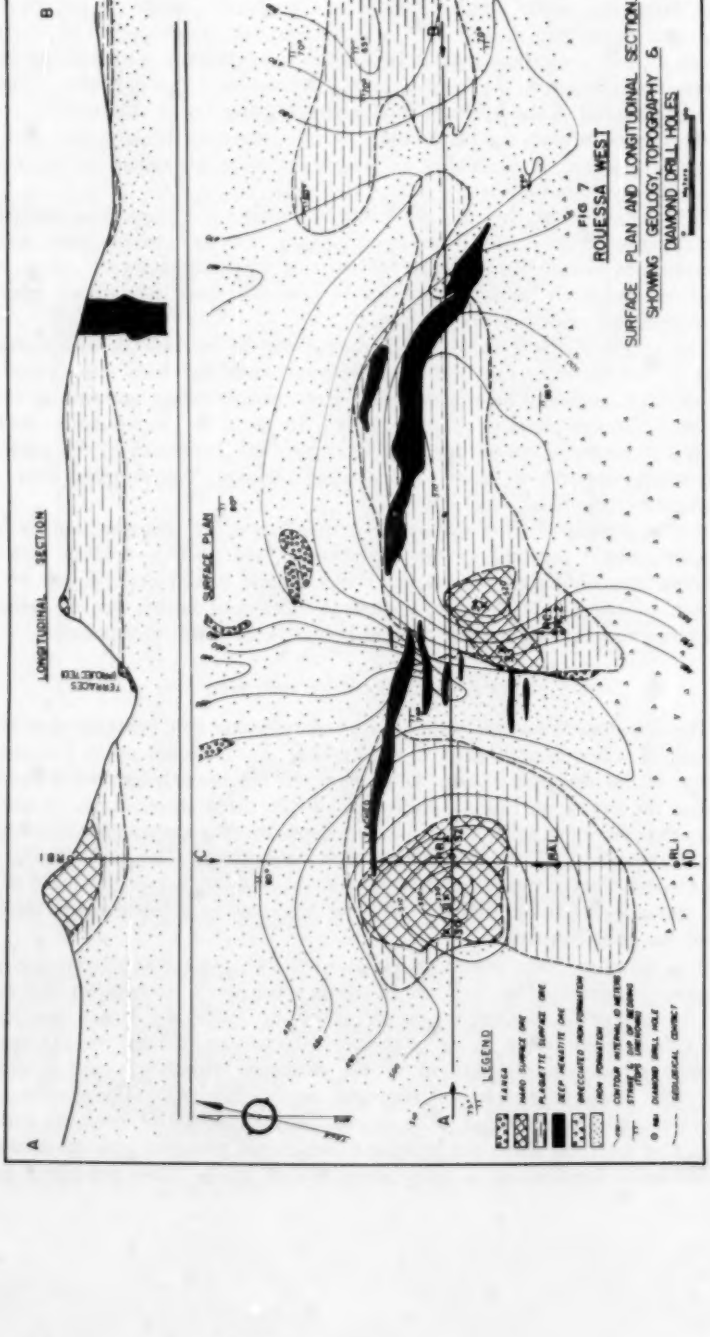
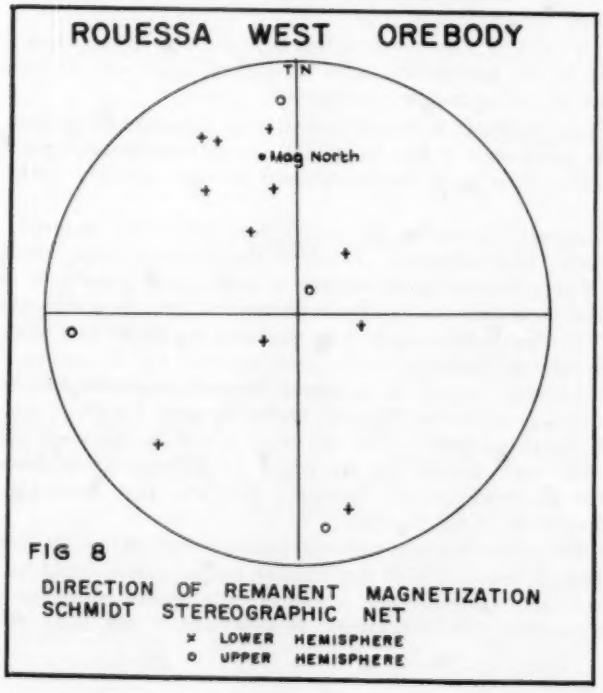


FIG 7
ROUESSA WEST
SURFACE PLAN AND LONGITUDINAL SECTION
SHOWING GEOLOGY, TOPOGRAPHY & DIAMOND DRILL HOLES
LONGITUDINAL SECTION
SURFACE PLAN
LEGEND
HANGING WALL SURFACE DIKE
FOOTWALL SURFACE DIKE
DEEP HEMATITE DIKE
UNACCOUNTED IRON FORMATION
IRON FORMATION
CONTOUR INTERVAL 5 METRE
DIAMOND DRILL HOLE
GEOLOGICAL CONTACT

TABLE 2

MAGNETIC MEASUREMENTS ON ROUESSA WEST ORE BODY

| Sample No. | Declination | Inclination |
|---|---|---|
| 30-1 | 92 | -68 |
| -2 | 118 | -75 |
| -3 | 25 | -80 |
| 31-1 | 265 | -15 |
| -2 | 253 | -38 |
| 32-1 | 338 | -28 |
| -2 | 335 | -26 |
| -3 | 331 | -60 |
| -4 | 36 | -62 |
| 33-1 | 325 | -40 |
| -2 | 253 | -24 |
| -3 | 353 | -26 |
| -4 | 170 | -18 |
| 34-1 | 354 | -48 |
| -2 | 358 | -13 |
| -3 | 158 | -16 |



netism in the biscuits and more massive pieces of hematite would tend to slump into a more-or-less random orientation.

The fairly prominent magnetic direction parallel to the present earth's field may be caused by the solution and redeposition of the soft hematite into a hard crust, which would have to take place in fairly modern times and perhaps during the Pleistocene when the climate in this part of Africa was less arid than it is today.

Concentrations of the Rouessa type are predominantly surficial and prospecting for additional ore is straightforward. The small concordant lenses of hematite are believed to be of the F'Derik type. Although they are relatively small in size and would not form an ore body on their own, the presence of these lenses might explain why surface ore was able to accumulate at Rouessa as opposed to other parts of the iron formation where these lenses do not occur.

## CONCLUSIONS

The following conclusions have been reached:

(a) Iron formations and particularly hematite ore bodies within iron formations are good subjects for study of remanent magnetism. The magnetic intensity is low, and the present investigation of a folded formation indicates that the stability is good.

(b) It is expected that work of this type could give important geological information on the age of deposition of iron minerals such as hematite and on the time of folding and metamorphism.

(c) Where sufficient information about the remanent magnetism is available in any given area it may be possible to determine the top direction of sediments that have been metamorphosed and the primary structures obliterated.

(d) Syngenetic hematite ore would have the same magnetic direction as the enclosing iron formation. If folded, the direction of magnetism on each limb should be separated by the degree of folding and if the beds were "unfolded" the direction of magnetism on either limb, if stable, should fall together. The inclination of the pole positions should be that of the earth's field at the time of deposition.

(e) If hematite formed in structural areas by metamorphic or igneous processes then, provided the hematite cooled through the Curie temperature, the magnetism in all parts of the ore body would be the same. However, magnetic directions within the ore could be different from the magnetic direction in surrounding iron formation that was not elevated above the Curie temperature of its minerals.

(f) If there were solution and redeposition of hematite, to form an ore concentration, a new direction of remanent magnetization could be acquired by this material. This direction would be independent of the magnetic direction in material already present. If redeposition took place at the time

of folding or later, the magnetic directions on the two limbs of a fold should be the same and reflect the direction of the earth's field at the time of redeposition.

(g) If the gangue minerals are leached by surface weathering, the hematite may concentrate as a residual deposit. Any original magnetic direction in the hematite would probably take on a random orientation by the slumping of the hematite. However, if there is sufficient solution and redeposition of hematite so that a hard crust is formed over the soft ore, then a remanent magnetism might develop that is aligned with the earth's field at the time of the redeposition.

(h) The hematite ores at Fort Gouraud are believed to have had a complex and multiple origin which might be summarized as follows:

| Ore forming process | Example | Present physical condition of ore | Approximate age |
|---|---|---|---|
| i. Formed during sedimentation or leaching of flat-lying iron formation. | F'Derik main zone. Small lenses Rouessa. | Hard | Precambrian as 1.F now overlain by flat Cambrian also magnetic poles had steep inclinations when formed. |
| ii. Formed during folding and metamorphism. | Some of the ore in the F'Derik main zone. | Hard | Precambrian. |
| iii. Formed by surface leaching. | Rouessa main ore zone. F'Derik surface ore. | Hard and soft. | Post folding some at least fairly modern. |
| iv. Formed by mechanical weathering and erosion of existing ores (talus). | F'Derik. | Hard | At least two ages of talus one being modern. |

ACKNOWLEDGMENTS

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UNIVERSITY OF TORONTO,
TORONTO, ONTARIO,
*June 11, 1961*

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# THE IRON CONTENT OF SPHALERITE FROM THE CENTRAL DISTRICT, NEW MEXICO AND THE BINGHAM DISTRICT, UTAH

ARTHUR W. ROSE

ABSTRACT

In conjunction with a study of trace elements, about 200 sphalerites from the Central district, New Mexico and the Bingham district, Utah have been analyzed for iron by an X-ray fluorescence method. The FeS content ranges from 4.2 to 18.5% at Hanover in the Central district and from 0.84 to 9.4% in the southwestern part of the same district. At Bingham the range is from 0.2 to 16.7%. Pyrite is ubiquitous in both districts, and pyrrhotite is known. Sphalerite with pyrrhotite averages higher in iron content than sphalerite with pyrite. A poorly-developed lateral zoning away from igneous centers is present in the Central district, and poorly-developed vertical zoning was found in one structure at Bingham.

Ten analyses of a zoned crystal from the Central district show a range from 4.3 to 12.1% FeS, and sharp boundaries exist between the zones, indicating that equilibrium was not achieved at temperatures of at least 425° C. Several other inhomogeneous sphalerites were found in both districts, and lack of equilibrium is believed to be a major difficulty in the determination of temperature from the iron content of sphalerite. Other difficulties include the effect of sulfur pressure, the total pressure correction, and exsolution effects. Because all these factors probably result in a lower iron content than is indicated by the FeS-ZnS solvus for one atmosphere, it is suggested that temperatures derived from the iron content of sphalerite be regarded as minimum temperatures unless convincing evidence can be presented on the possible complications.

On this basis, temperatures reached at least 600° C at Hanover in the Central district, at least 350° C at the Groundhog mine in the same district, at least 650° C at the Cleveland mine in the nearby Pinos Altos district, and at least 550° C in the Bingham district. At the Cleveland mine, higher than normal $CO_2$ pressures seem necessary to explain the lack of reaction between quartz and carbonates, if the sphalerite temperatures are correct.

## INTRODUCTION

In 1953, Kullerud (16) published an experimentally determined phase diagram for the system FeS–ZnS and proposed that the amount of iron in sphalerite, under certain conditions, could be used as a geological thermometer. Since then, an increasing amount of interest has been shown in the method, and many papers on the subject have been published. The present study was undertaken as part of an investigation of the trace element content of sphalerite and chalcopyrite from the Central mining district, New Mexico, and the Bingham mining district, Utah (22). The samples were not col-

lected with the specific intent of studying their iron content, but they represent a reasonably good cross-section of the two districts and therefore seem worthy of discussion.

## ANALYTICAL METHOD

The analytical method will be described in a separate paper (7), but a brief summary follows. The method uses X-ray fluorescence spectrography and requires only three to ten milligrams of sphalerite.

The samples for analysis are drilled from a polished section with a dental drill, choosing areas that are free of contaminating minerals and noting any minerals that might be possible contaminants. Impurities of iron-bearing minerals are believed to amount to less than one percent except in a few specimens containing chalcopyrite and/or pyrrhotite blebs distributed throughout the sphalerite. These blebs may have been formed by exsolution, but in any case they are not believed to exceed about two percent, based on megascopic estimates and a few X-ray fluorescence analyses for copper.

From three to ten milligrams of powdered sample is placed in a shallow well drilled in a lucite holder and inserted in the X-ray beam. The iron and zinc peak intensities are measured and background counted adjacent to each peak. Standards consist of nine sphalerites analyzed by wet chemical methods for iron and zinc by E. Godijn and the writer. The FeS content of the standards ranges from 0.33 to 18.17 percent. The Zn/Fe intensity ratio is plotted against the ZnS/FeS weight ratio to give a nearly straight calibration curve. The percent FeS is found from the relation $\%FeS = 100/(1 + ZnS/FeS)$. This relation neglects the small amounts of manganese, cadmium, and mineral impurities present. However, the small amounts of manganese and cadmium in the samples, as determined by optical spectrography, would have a negligible effect on the percent FeS found, and the iron-bearing impurities pyrrhotite and chalcopyrite amount to a maximum of 2 percent. The method ignores iron-free and zinc-free impurities, giving the composition of the sphalerite alone in this case.

Repeated analysis of the standards gave a coefficient of variation of 3 percent, except in the range below 1% FeS where it increases to 5–10 percent.

## FACTORS INFLUENCING THE IRON CONTENT OF SPHALERITE

The determination of temperature by the iron content of sphalerite, as suggested by Kullerud (16), requires equilibrium between sphalerite and FeS at the time for which the temperature is to be determined. For the method to be used on natural sphalerites, the iron content must have remained unchanged during cooling and other events subsequent to this time, or if exsolution can be proven, the exsolved material must have remained within the volume of material analyzed. At equilibrium, the sphalerite has the composition of the solvus of the FeS–ZnS system for that temperature. The solvus as used by the writer is shown in Figure 1, based on data by Kullerud (16). It should be noted that although deposition temperatures

below 400° C are quite common geologically, the solvus determined by Kullerud for these temperatures is extrapolated, and the extrapolation method is of unknown accuracy.

In addition to temperature, the solubility of iron in sphalerite depends slightly on total pressure, because of the volume increase in the formation of iron-bearing sphalerite from ZnS and FeS. Kullerud calculated a pressure correction of about 25° C per 1,000 bars total pressure, and showed experimentally that pressure decreased the solubility of FeS in ZnS.

Another very significant factor in determining the iron content of natural sphalerites is the partial pressure or chemical potential of sulfur in the system at the time of equilibrium. Pyrrhotite in nature is slightly deficient in iron relative to the composition FeS. Kullerud (16) found almost no effect on the iron content of sphalerite as long as some pyrrhotite was present in the system. More recent experiments (3) indicate a large effect of sulfur pressure when pyrite occurs with the sphalerite and no pyrrhotite is present. According to Barton and Kullerud (3), "Temperatures estimated using the

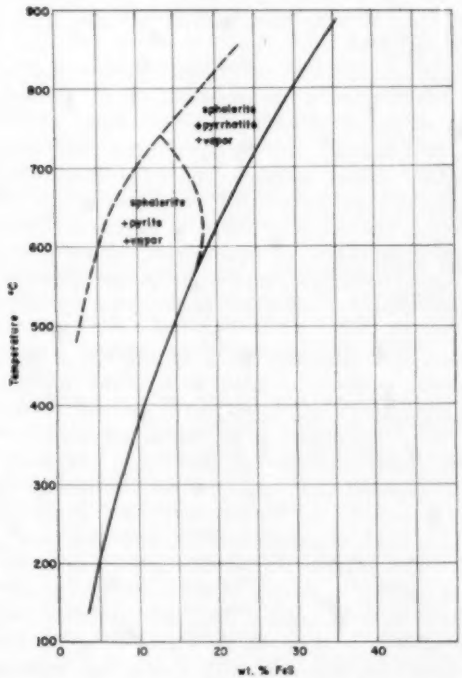


FIG. 1. Solvus of the system ZnS–FeS, as used in this study, after Kullerud. Dashed lines indicate limits of solid solution in the system ZnS–FeS–S, after Barton and Kullerud.

FeS–ZnS data for sphalerite in equilibrium with pyrite can be as much as 300° C too low." In addition, Barton and Kullerud (4) have found that sphalerite in equilibrium with non-stoichiometric pyrrhotite above about 600° C can also contain considerably less iron than indicated by the FeS–ZnS solvus. The dashed lines in Figure 1 indicate the approximate limits of composition of sphalerite in equilibrium with pyrrhotite or pyrite plus liquid and vapor. It can be seen that a sphalerite containing 10% FeS and formed in equilibrium with pyrite, might be deposited at any temperature from about 375° to about 675° C. Similar ranges of temperature exist for sphalerite in equilibrium with pyrrhotite above about 600° C.

Another possible complication is the effect of solid solution between chalcopyrite and sphalerite. This solid solution has been shown to occur by Buerger (5) and Donnay and Kullerud (8). According to the latter reference, almost 10 percent chalcopyrite dissolves in ZnS at 600° C. If appreciable amounts of chalcopyrite are dissolved in sphalerite, or were dissolved at the temperature of deposition but have since exsolved, some effect on the iron content is likely. The exact effect depends on the amount of solid solution and on the cooling history of the specimen. The common occurrence of chalcopyrite blebs in sphalerite indicates that this phenomenon may be of some significance.

Lack of equilibrium and lack of criteria for demonstrating equilibrium are also important difficulties in the application of the method. Evidence for disequilibrium, in the form of strong zoning within sphalerite crystals apparently formed at relatively high temperatures, was found in both the Central and Bingham mining districts, and is probably common elsewhere. In addition, much evidence shows that ore minerals are usually deposited in a sequence (19). Therefore it cannot be merely assumed that sphalerite and pyrrhotite were deposited in equilibrium. Even if the two minerals were deposited simultaneously, equilibrium is not guaranteed; deposition might have been so rapid that there was not time for equilibrium distribution of components.

If the minerals were deposited in a sequence, for instance pyrrhotite followed by sphalerite, equilibrium might be achieved in one of several ways. Transfer of iron from pyrrhotite to sphalerite or vice versa might occur by diffusion within the solid phases after deposition of both minerals. The evidence of zoned crystals mentioned above and discussed later indicates that this process is probably not very effective. A second alternative is the partial solution of pyrrhotite by the ore fluid during the deposition of the sphalerite to yield a fluid saturated in FeS, and thus perhaps a sphalerite saturated in FeS. The effectiveness of this process depends on the relative rates of solution of pyrrhotite, mixing of fluid, and precipitation of sphalerite, but there appears to be no guarantee that an equilibrium sphalerite will result. A third possibility is that the fluid depositing sphalerite arrived at the site of deposition just saturated with FeS. Any evidence for replacement of pyrrhotite by sphalerite disproves this possibility. In such cases, it is clear that the hydrothermal fluid was not in equilibrium with the pyrrho-

tite because the pyrrhotite was dissolving in either the fluid phase or the sphalerite or both and was thus unstable.

In summary, evidence for equilibrium is generally lacking whereas evidence for disequilibrium is not uncommon, especially in deposits of moderate to low temperatures of deposition. The approach of testing for consistent temperatures by two or more methods of thermometry as suggested by Kullerud (17) has promise for appropriate mineral assemblages.

Available data on reaction rates between solid sphalerite and pyrrhotite are contradictory. Kullerud obtained equilibrium by solution at 500° C in about a year. However, he found difficulty in explaining the lack of exsolution in sphalerites of high iron content, such as those at Broken Hill, Australia, which show relatively little exsolved pyrrhotite, but would be expected to show considerable pyrrhotite if they cooled from the inferred temperature of 600°+ C over a period of years, as seems reasonable. Similar problems have been encountered and discussed by Edwards and Lyon (10) and Lyon (20). Kullerud (16) suggests that impurities may be responsible for the slow exsolution of natural sphalerites, but this interpretation does not apply to the experiments of Lyon (20) using natural materials.

Fryklund and Fletcher (11) stated that sphalerite near a post-ore dike developed iron-rich rims where it was in contact with iron minerals (magnetite, pyrite, and siderite) presumably because of the high temperature reached during the intrusion and crystallization of the dike. On the other hand, sphalerite in most of the ore showed no such rims, and they concluded that equilibrium must have existed at the time of deposition, or a similar zoning would have existed in all the sphalerite. This argument may not be valid because of the difference in reaction rates at various temperatures and because the iron minerals involved are not pyrrhotite. The addition of iron from pyrite, as apparently happened adjacent to the dikes, involves a loss of sulfur from the mineral assemblage. This loss might occur readily under the conditions of dike intrusion, but might be prevented during deposition of the ore by a much higher sulfur pressure in the ore fluid. Reaction with magnetite or siderite involves similar complications with regard to reaction products.

Barton and Kullerud (3) have suggested that the temperatures obtained from sphalerite occurring with pyrite, using the FeS–ZnS solvus, should be interpreted as minimum temperatures, that is, the temperature was equal to or higher than the value found. To the writer, it seems reasonable to regard all sphalerite temperatures in a similar light, unless definite evidence regarding all the complicating factors is at hand. For instance, use of a pressure correction raises the temperature above the value found from the solvus for one atmosphere. Exsolution of pyrrhotite, equilibrium with pyrite, or equilibrium with pyrrhotite above about 600° C have a similar effect. If equilibrium did not exist, a sphalerite undersaturated in iron seems more likely to result than an oversaturated one. Therefore, because of the lack of definite criteria for equilibrium and the difficulty of making a pressure correction, a minimum temperature interpretation seems realistic for all but

a few sphalerites at the present stage of knowledge. Nevertheless, it should be recognized that complications like the formation of a supersaturated solid solution, impure samples, or perhaps solid solution of chalcopyrite may invalidate this minimum temperature interpretation.

## SPHALERITES FROM THE CENTRAL MINING DISTRICT, NEW MEXICO

The Central mining district of New Mexico is best known for its production of copper from the Chino open pit mine at Santa Rita, but the district, encompassing the area around the towns of Fierro, Hanover, Santa Rita, Vanadium, Bayard and Central, also contains notable deposits of zinc and iron, and smaller amounts of lead, silver and gold. Geological features of the district have been discussed by Spencer and Paige (25), Schmitt (23, 24), Lasky and Hoagland (18), Hernon, Jones, and Moore (13) and others.

In late Cretaceous or early Tertiary time, several varieties of intermediate to acid igneous rock were emplaced in Paleozoic and Mesozoic sediments, and the mineral deposits were formed. Ore deposition occurred following the intrusion of granodiorite stocks and apparently overlapped a period of intrusion of quartz monzonite dikes. Considerable faulting and brecciation occurred during the period of mineralization. Several varieties of monzonitic rocks were intruded during a post-ore period of igneous activity, and later Miocene (?) vulcanism produced a number of dikes and blanketed the area with flows.

The zinc deposits of the Empire Zinc and Pewabic mines around the south end of the Hanover-Fierro stock at Hanover (Fig. 2) are mainly replacements of Mississippian and Pennsylvanian limestones, principally the Mississippian Hanover limestone. The ore bodies range in shape from relatively narrow tabular replacements adjacent to faults, fissures and dike contracts to wider more irregular replacement bodies roughly paralleling the contact. The principal gangue minerals are andradite, hedenbergite and magnetite with smaller amounts of quartz, carbonates, ilvaite, epidote and other silicates. Pyrite is present in most specimens of ore and pyrrhotite is present in some. Minor amounts of chalcopyrite and galena are present. The sphalerite commonly contains sparse to abundant rounded blebs of chalcopyrite and in a few cases blebs of pyrrhotite. Pyrite occurs as cubes partly replaced by sphalerite, chalcopyrite and galena, and also as tiny veinlets of fine-grained late pyrite cutting sphalerite, chalcopyrite and galena. Much pyrite occurs as anhedral grains that could not be definitely classified in age relative to sphalerite. This pyrite may have been deposited before, contemporaneously with or after sphalerite.

Pyrrhotite is more common at the Pewabic mine than at the Empire Zinc mine. In the polished sections studied, it occurs as small ragged grains and as blebs in sphalerite. It is also inferred to have been present in other specimens now showing areas of lamellar or colloform marcasite that is interpreted to have replaced pyrrhotite in a few polished sections. The grains of pyrrhotite appear to have preceded sphalerite, but the blebs in sphalerite may result either from exsolution or later replacement.

In the Oswaldo No. 1, Oswaldo No. 2 and Kearney mines between the Hanover-Fierro and Santa Rita stocks, there occur zinc ores of mineralogical character similar to these at Hanover, but generally with more definite structural control.

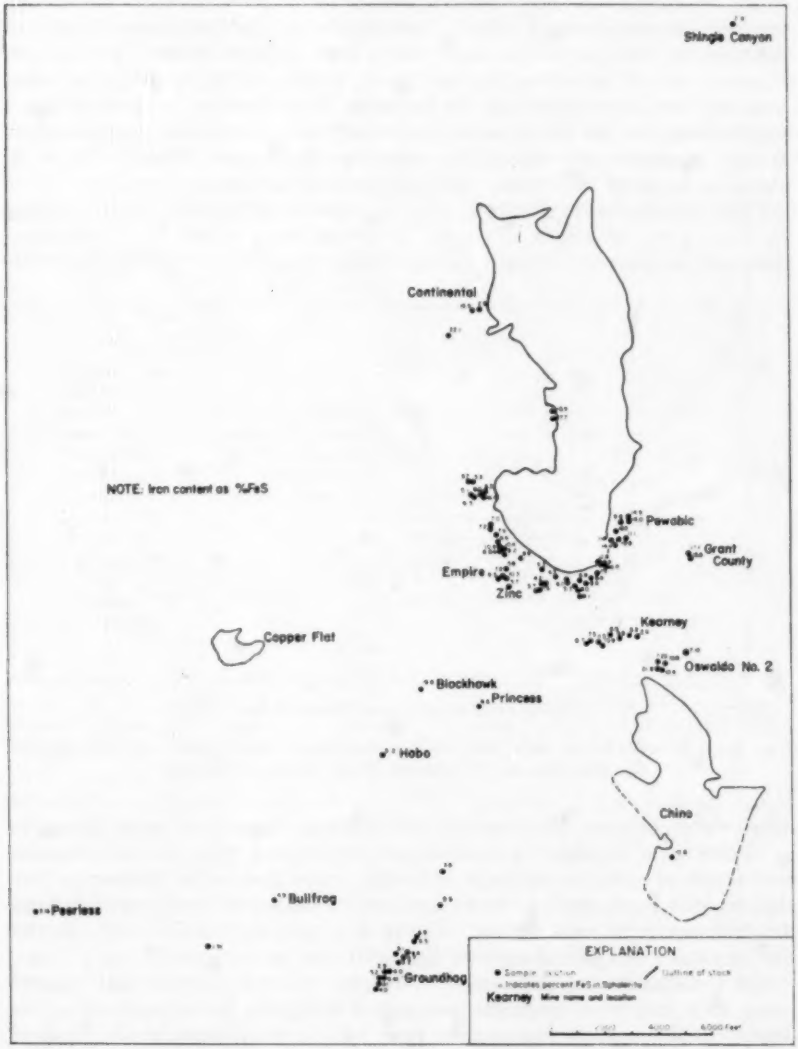


FIG. 2. Map of the Central mining district, New Mexico, showing stocks, mines, sphalerite sample locations and FeS contents.

The Groundhog mine in the southwestern corner of the district was originally developed on argentiferous zinc-lead-copper ore occurring as veins in quartz diorite and Mesozoic sediments. Deeper ore mined more recently has been dominantly of the limestone replacement type, similar in character to the ore mined near Hanover, but with different proportions of the various ore and gangue minerals. Both types of ore at the Groundhog mine are localized by the Groundhog fault zone of the northeast trend. The gangue minerals include hedenbergite, specularite, pyrite, calcite, quartz, clay minerals and chlorite. Sphalerite is the most important ore mineral. Galena and chalcopyrite are economically important but appear to be less abundant in the limestone replacement ore than in the fissure fillings. Pyrite is abundant in many specimens. No pyrrhotite is reported.

The deposits in the Bullfrog mine are similar to those in the Groundhog

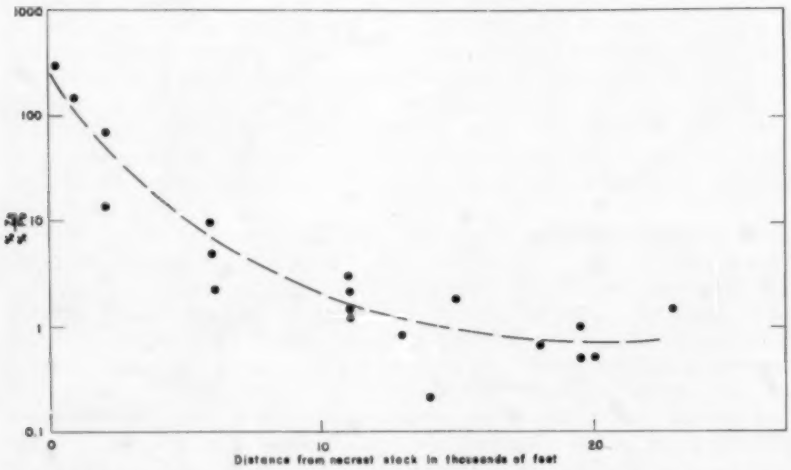


FIG. 3. Plot of Pb/Zn ratio of ore samples from the Central mining district against distance to nearest stock, showing zoning.

mine. Ores between Hanover and the Bullfrog-Groundhog area appear to be intermediate between the Hanover and Groundhog types in character and abundance of gangue minerals, lead-zinc ratio and other features. The zinc deposits in the main part of the district thus appear to be zoned around the Hanover area as a center. Figure 3 shows a gradual change in the Pb/Zn ratio when plotted against distance to the nearest stock.

At the Continental mine, west of Fierro, sphalerite occurs with chalcopyrite in a gangue of magnetite and garnet along the Barringer fault. The deposit is of limestone replacement type. Farther northeast at the Shingle Canyon mine, sphalerite replaces limestone and shale, also along the Barringer fault.

West of the Bullfrog and Groundhog mines there are several small mines producing silver, lead and zinc from veins. The largest of these is the Peerless mine at Central. In addition, zinc has been produced from limestone replacement deposits at Copper Flats (21), where a small plug of quartz monzonite intrudes upper Paleozoic sediments.

The inferred sequence of deposition in ores from the district, based on polished section work by the writer and data from the literature, is indicated in Figure 4.

Evidence on the problem of post-depositional equilibration and on the effect of temperature and ore fluid composition is contributed by a large crystal of sphalerite from a vug at Hanover (Fig. 5). The crystal is about 3 cm in diameter and the cut surface shows a number of sharp zonal color changes caused mainly by changes in the amount of iron present in the sphalerite. The bands were formed at successive stages in the growth of the crystal. Quartz, siderite, manganiferous (?) calcite and chalcopyrite were

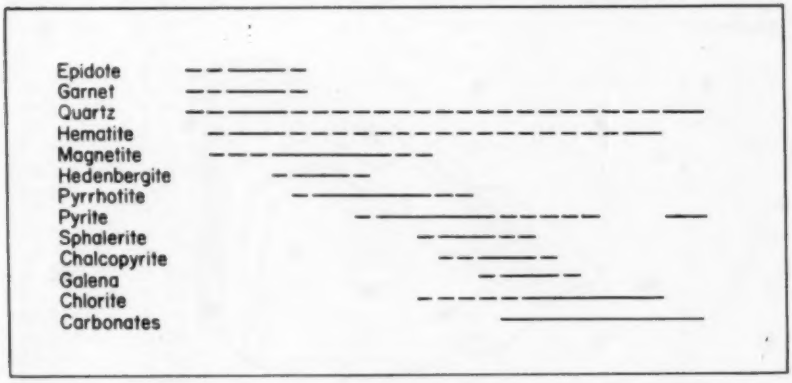


FIG. 4. Generalized paragenesis of minerals in ore deposits of the Central mining district after Rose (22), Schmitt (24), and Lasky and Hoagland (18). Dashes indicate uncertainty or differences between parts of the district.

deposited as euhedral crystals on top of the sphalerite in the sequence listed. Marcasite is present at the base of the crystal. Sparse chalcopyrite and galena occur within it, but may be of replacement origin.

The analyses of ten samples drilled from different color bands in the sphalerite are listed in Table 1. The FeS content ranges from 4.35 to 12.1 percent. Significantly, a zone with an FeS content of 12.1 percent (indicating a temperature of at least 425° C) overlies a zone with 5.4 percent FeS (at least 230° C). The zones are separated along a line that is sharp under examination by a hand lens. Several occurrences of sphalerite from Bingham also showed rims of high iron content surrounding areas of relatively low iron content. This arrangement of zones indicates (1) that the time during which the temperature of at least 425° C was maintained was

TABLE 1

| Sample number | % FeS | |
|---|---|---|
| CEN-531-1 | 4.35 | Surface of crystal |
| CEN-531-2 | 7.57 | |
| CEN-531-3 | 12.1 | |
| CEN-531-4 | 12.1 | |
| CEN-531-5 | 7.70 | |
| CEN-531-6 | 7.05 | |
| CEN-531-7 | 5.41 | |
| CEN-531-8 | 8.85 | |
| CEN-531-9 | 10.65 | |
| CEN-531-10 | 7.63 | Base of crystal |

so brief that no perceptible diffusion from one zone to the next took place, or (2) that the reaction rates found by Kullerud (16) are considerably too fast, or (3) that an unknown complication makes the iron content invalid even as an index of minimum temperature. The writer prefers to believe one of the first two alternatives, but all three alternatives cast doubt on the ability of reactions between or within solids to establish equilibrium in natural sphalerite.

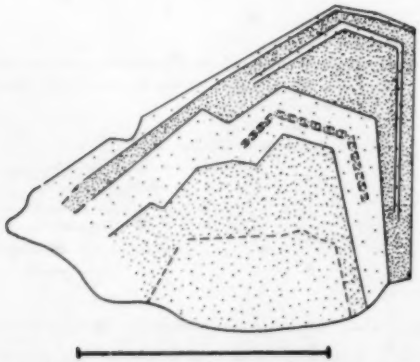

FIG. 5. Sketch of cut surface of sphalerite single crystal from Hanover, New Mexico, showing color zoning. Bar indicates one inch.

The FeS content of sphalerite from the Central mining district is listed in Table 2a and shown on Figure 2. As indicated in the table, only ten of the analyzed specimens contained pyrrhotite. The numbers and locations are given in Table 2a. The FeS content of these specimens ranges from 9.1 to 19.8% (CEN-520 and 243 contained about 2% pyrite or pyrrhotite and the iron content of the sphalerite itself is thus not as high as the analysis would indicate). The minimum temperatures indicated by the values range from 350° C to slightly over 600° C, with most values in the 500–600° C range. Pyrrhotite is present only as a few blebs in CEN-193, 149, 160, and 578, and as blebs plus a few discrete grains in CEN-151, 152, 162, 520, 243, and BH-18. Because pyrrhotite is present with these sphalerites, and because of their generally high FeS content, these ten samples probably give

TABLE 2a

Iron Content of Sphalerite from the Central and Pinos Altos Mining Districts, New Mexico

| Sample No. | FeS(%) | Other Minerals Present: Pyrite | Pyrrhotite | Chalcopyrite | Locality | |
|---|---|---|---|---|---|---|
| CEN-168-3 | 16.5 | x | | x* | Empire Zinc Mine | |
| CEN-168-5 | 16.6 | x | | x* | " | " |
| CEN-171 | 15.9 | | | | " | " |
| CEN-174 | 6.23 | x | | | " | " |
| CEN-177 | 5.65 | x | | x | " | " |
| CEN-178 | 5.33 | x | | x | " | " |
| CEN-180 | 7.25 | x | | x* | " | " |
| CEN-183 | 7.00 | x | | | " | " |
| CEN-188 | 7.87 | x | | x* | " | " |
| CEN-189 | 6.63 | x | | x* | " | " |
| CEN-190 | 10.86 | x | | x* | " | " |
| CEN-193-1 | 9.08 | x | | x* | " | " |
| CEN-193-2 | 11.6 | x | x | x* | " | " |
| CEN-194 | 6.06 | x | | x* | " | " |
| CEN-195 | 8.20 | x | | x* | " | " |
| CEN-197 | 5.83 | | | x* | " | " |
| CEN-198 | 8.78 | x | | x* | " | " |
| CEN-199 | 5.19 | x | | x* | " | " |
| CEN-203 | 6.32 | x | | x* | " | " |
| CEN-207 | 6.00 | x | | | " | " |
| CEN-208 | 6.73 | | | x* | " | " |
| CEN-209 | 8.93 | x | | x* | " | " |
| CEN-211 | 6.85 | x | | x* | " | " |
| CEN-213 | 12.8 | x | | x* | " | " |
| CEN-213A | 7.93 | x | | x* | " | " |
| CEN-214 | 8.61 | | | | " | " |
| CEN-216 | 8.78 | x | x | x* | " | " |
| CEN-218 | 5.29 | x | | x* | " | " |
| CEN-219 | 5.65 | x | x | x* | " | " |
| CEN-220 | 7.64 | | | x* | " | " |
| CEN-222 | 4.21 | x | | x* | " | " |
| CEN-227 | 9.47 | x | | x* | " | " |
| CEN-229-1 | 10.3 | x | | x* | " | " |
| CEN-230 | 5.71 | x | | x* | " | " |
| CEN-233 | 5.05 | | | x* | " | " |
| CEN-533 | 7.35 | x | | x* | " | " |
| CEN-534 | 5.88 | x | | x* | " | " |
| CEN-535 | 7.58 | | | x* | " | " |
| CEN-536 | 7.25 | x | | x* | " | " |
| CEN-537 | 7.25 | x | | x* | " | " |
| CEN-538 | 5.32 | x | | x* | " | " |
| CEN-538A-1 | 7.70 | x | | x* | " | " |
| CEN-538A-2 | 7.05 | x | | x* | " | " |
| CEN-538A-3 | 8.85 | x | | x* | " | " |
| CEN-541 | 5.55 | x | | | " | " |
| CEN-542 | 5.16 | | | x* | " | " |

a closer estimate of the maximum temperature of deposition of sphalerite in the area than the many remaining samples. However, because of the possibility that sphalerite was deposited over a range in temperature, the

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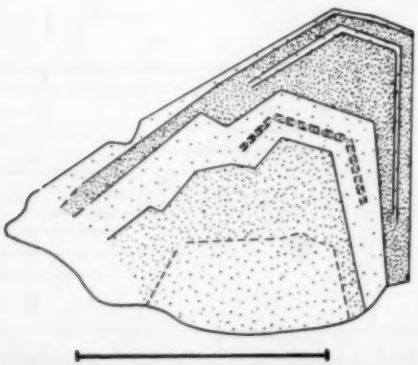

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| Sample No. | FeS(%) | Other Minerals Present: Pyrite | Pyrrhotite | Chalcopyrite | Locality | |
|---|---|---|---|---|---|---|
| CEN-168-3 | 16.5 | x | | x* | Empire Zinc Mine | |
| CEN-168-5 | 16.6 | x | | x* | " | " |
| CEN-171 | 15.9 | | | | " | " |
| CEN-174 | 6.23 | x | | | " | " |
| CEN-177 | 5.65 | x | | x | " | " |
| CEN-178 | 5.33 | x | | x | " | " |
| CEN-180 | 7.25 | x | | x* | " | " |
| CEN-183 | 7.00 | x | | | " | " |
| CEN-188 | 7.87 | x | | x* | " | " |
| CEN-189 | 6.63 | x | | x* | " | " |
| CEN-190 | 10.86 | x | | x* | " | " |
| CEN-193-1 | 9.08 | x | | x* | " | " |
| CEN-193-2 | 11.6 | x | x | x* | " | " |
| CEN-194 | 6.06 | x | | x* | " | " |
| CEN-195 | 8.20 | x | | x* | " | " |
| CEN-197 | 5.88 | | | x* | " | " |
| CEN-198 | 8.78 | x | | x* | " | " |
| CEN-199 | 5.19 | x | | x* | " | " |
| CEN-203 | 6.32 | x | | x* | " | " |
| CEN-207 | 6.00 | x | | | " | " |
| CEN-208 | 6.73 | | | x* | " | " |
| CEN-209 | 8.93 | x | | x* | " | " |
| CEN-211 | 6.85 | x | | x* | " | " |
| CEN-213 | 12.8 | x | | x* | " | " |
| CEN-213A | 7.93 | x | | x* | " | " |
| CEN-214 | 8.61 | | | | " | " |
| CEN-216 | 8.78 | x | x | x* | " | " |
| CEN-218 | 5.29 | x | | x* | " | " |
| CEN-219 | 5.65 | x | x | x* | " | " |
| CEN-220 | 7.64 | | | x* | " | " |
| CEN-222 | 4.21 | x | | x* | " | " |
| CEN-227 | 9.47 | x | | x* | " | " |
| CEN-229-1 | 10.3 | x | | x* | " | " |
| CEN-230 | 5.71 | x | | x* | " | " |
| CEN-233 | 5.05 | | | x* | " | " |
| CEN-533 | 7.35 | x | | x* | " | " |
| CEN-534 | 5.88 | x | | x* | " | " |
| CEN-535 | 7.58 | | | x* | " | " |
| CEN-536 | 7.25 | x | | x* | " | " |
| CEN-537 | 7.25 | x | | x* | " | " |
| CEN-538 | 5.32 | x | | x* | " | " |
| CEN-538A-1 | 7.70 | x | | x* | " | " |
| CEN-538A-2 | 7.05 | x | | x* | " | " |
| CEN-538A-3 | 8.85 | x | | x* | " | " |
| CEN-541 | 5.55 | x | | | " | " |
| CEN-542 | 5.16 | | | x* | " | " |

a closer estimate of the maximum temperature of deposition of sphalerite in the area than the many remaining samples. However, because of the possibility that sphalerite was deposited over a range in temperature, the

| Sample No. | FeS(%) | Other Minerals Present: Pyrite | Pyrrhotite | Chalcopyrite | Locality |
|---|---|---|---|---|---|
| CEN-543 | 7.00 | | | x* | Empire Zinc Mine |
| CEN-544 | 8.05 | | | x* | " " |
| CEN-545 | 6.76 | x | | x* | " " |
| CEN-531A | 17.9 | x | | x* | " " |
| CEN-530 | 8.00 | x | | x* | " " |
| HAN-6 | 13.9 | x | | x* | " " |
| HAN-5 | 9.52 | x | | x* | " " |
| CEN-149 | 18.5 | | x | x | Pewabic Mine |
| CEN-150 | 9.80 | x | | x | " " |
| CEN-151 | 14.9 | | x | x | " " |
| CEN-152-1 | 17.9 | x | x | | " " |
| CEN-152-2 | 18.6 | x | x | | " " |
| CEN-152-3 | 17.8 | x | x | | " " |
| CEN-152-4 | 17.9 | x | x | | " " |
| CEN-153 | 11.1 | | | x | " " |
| CEN-158 | 10.9 | | | x | " " |
| PEW-8 | 11.4 | | | | " " |
| PEW-9 | 14.4 | | | | " " |
| CEN-236 | 6.70 | x | | x* | Kearney Mine |
| CEN-237 | 7.25 | x | | x | " " |
| CEN-238 | 6.80 | x | x | x* | " " |
| CEN-239 | 5.20 | | | | " " |
| CEN-240 | 5.89 | x | | x* | " " |
| CEN-241 | 6.45 | x | | x* | " " |
| CEN-577 | 11.3 | x | | x | " " |
| CEN-578 | 15.9 | x | x | x* | " " |
| CEN-518 | 7.10 | x | | x | Oswaldo #2 Mine |
| CEN-519 | 10.9 | x | | x* | " " |
| CEN-520 | 21.8 | x | | x | " " |
| CEN-521 | 10.4 | x | | x* | " " |
| CEN-522 | 7.25 | x | | x | " " |
| CEN-523 | 8.27 | x | | x* | " " |
| CEN-160 | 17.1 | x | x | x* | Grant County Mine |
| CEN-162 | 13.9 | x | x | x* | " " |
| CEN-255 | 6.90 | x | | x* | Groundhog Mine |
| CEN-256 | 6.46 | x | | x* | " " |
| CEN-257 | 9.42 | x | | x* | " " |
| CEN-264 | 6.41 | x | | x* | " " |
| CEN-265 | 5.22 | x | | x* | " " |
| CEN-266-1 | 2.74 | x | | x* | " " |
| CEN-266-2 | 3.03 | x | | x* | " " |
| CEN-266-3 | 3.54 | x | | x* | " " |
| CEN-266-4 | 3.20 | x | | x* | " " |
| CEN-266-5 | 3.40 | x | | x* | " " |
| CEN-267-1 | 0.34 | x | | x | " " |
| CEN-267-2 | 4.90 | x | | x | " " |
| CEN-266 | 4.00 | x | | x* | " " |
| CEN-269 | 3.12 | x | | x* | " " |
| CEN-271 | 4.00 | x | | x* | " " |
| CEN-274 | 4.72 | x | | x* | " " |
| CEN-3/1 | 3.78 | x | | x* | " " |
| G-3 | 3.06 | x | | x* | " " |
| SR-15 | 5.62 | x | | x | " " |

TABLE 2a (3)

| Sample No. | FeS(%) | Other Minerals Present: Pyrite | Pyrrhotite | Chalcopyrite | Locality |
|---|---|---|---|---|---|
| CEN-377 | 6.63 | x | | x | Princess Mine |
| BH-18 | 19.8 | | x | x* | Blackhawk Mine |
| CEN-348 | 3.16 | x | | x* | Hobo Mine |
| CEN-376 | 4.90 | x | | x | Bullfrog Mine |
| CEN-513 | 5.40 | x | | x* | " " |
| CEN-547 | 1.91 | x | | x* | Near Slate Mine |
| CEN-579 | 10.6 | x | | x | Chino Mine |
| CEN-359 | 17.7 | x | | | Republic Iron Mine |
| CEN-507 | 18.9 | x | | | " " |
| CEN-341 | 1.96 | x | | x* | Shingle Canyon Mine |
| CEN-243 | 22.1 | x | | x | Pearson Mine |
| CEN-358B | 2.86 | | | x* | Continental Mine |
| CEN-554 | 15.5 | x | | x* | " " |
| CEN-58 | 22.0 | x | x | x* | Cleveland Mine, Pinos Altos Dist. |
| CEN-60 | 24.1 | x | x | x* | " " " |
| CEN-61 | 22.8 | x | x | x* | " " " |
| CEN-62 | 19.9 | x | x | x* | " " " |
| CEN-63 | 21.2 | x | | x* | " " " |
| CEN-64 | 15.7 | x | | x* | " " " |
| CEN-285A | 22.2 | x | x | x* | " " " |
| CEN-286 | 19.3 | x | | x* | " " " |
| CEN-287 | 22.6 | x | x | x* | " " " |
| CEN-289A | 22.5 | x | x | x* | " " " |
| CEN-320 | 18.0 | x | x | x* | " " " |
| CEN-321 | 20.0 | x | | x | " " " |
| CEN-281 | 15.6 | x | | x* | Vein near highway to Pinos Altos |
| CEN-334-2 | 4.76 | | | x | Tiptop Shaft |
| CEN-337 | 0.37 | x | | x | Aztec shaft |
| CEN-338 | 1.0 | x | | x | " " |
| CEN-361 | 1.0 | x | | x* | Manhattan Mine |
| CEN-363-1 | 0.43 | | | | " " |

* = chalcopyrite as blebs in sphalerite.

minimum temperatures indicated by the ten samples cannot be taken as typical for sphalerite deposition in the district.

In addition, the possibility of lack of equilibrium with pyrrhotite is indicated by the differences between the two samples from the polished section of CEN-193 (9.1% & 11.6% FeS). A still darker variety of sphalerite is present in the polished section but contained too many chalcopyrite blebs for meaningful analysis. In any case, it is clear that equilibrium within the sphalerite, and therefore between the sphalerite and pyrrhotite, did not exist, suggesting the possibility of non-equilibrium in the other nine specimens. However, the considerably lower iron content in CEN-193 than in the rest of the pyrrhotite-bearing samples allows the interpretation that equilibrium was not reached at temperatures of 350–400° C, but was reached above 500° C. Minor differences in iron content are recorded in the four samples drilled from CEN-152, but the differences are within the analytical error.

TABLE 2b

Iron Content of Sphalerite from the Bingham Mining District, Utah

| Sample Number | FeS (%) | Other Minerals Present: Pyrite | Chalcopyrite | Locality |
|---|---|---|---|---|
| BIN-580-1 | 3.45 | x | x* | Holden fissure |
| BIN-580-2 | 0.55 | x | x* | " " |
| BIN-581 | 0.69 | x | x* | " " |
| BIN-583 | 0.51 | x | x* | Galena fissure |
| BIN-585 | 2.13 | x | x* | " " |
| BIN-586-1 | 0.33 | x | x* | " " |
| BIN-586-2 | 0.25 | x | x* | " " |
| BIN-586-3 | 0.93 | x | x* | " " |
| BIN-589-1 | 0.33 | x | x* | " " |
| BIN-589-2 | 0.68 | x | x* | " " |
| BIN-592 | 1.52 | x | x | " " |
| BIN-595-1 | 1.22 | x | x* | " " |
| BIN-595-2 | 6.75 | x | x* | " " |
| BIN-596-1 | 1.19 | x | x* | " " |
| BIN-596-2 | 1.25 | x | x* | " " |
| BIN-596-3 | 0.25 | x | x* | " " |
| BIN-596-4 | 2.06 | x | x* | " " |
| BIN-596-5 | 0.38 | x | x* | " " |
| BIN-596-6 | 0.25 | x | x* | " " |
| BIN-596-7 | 4.88 | x | x* | " " |
| BIN-596-8 | 0.61 | x | x* | " " |
| BIN-599 | 1.61 | x | x | " " |
| BIN-600 | 0.54 | x | x | " " |
| BIN-577 | 2.26 | x | x | Highland limestone |
| BIN-579 | 0.99 | x | x | " " |
| BIN-617 | 8.50 | x | x* | B limestone |
| BIN-625 | 8.55 | x | x* | " " |
| BIN-626 | 12.4 | x | x* | " " |
| BIN-646 | 14.5 | x | x* | " " |
| BIN-647 | 12.8 | x | x* | " " |
| BIN-654 | 13.8 | x | x* | " " |
| BIN-655 | 14.0 | x | x* | " " |
| BIN-671-1 | 8.27 | x | x* | " " |
| BIN-671-2 | 5.55 | x | x* | " " |
| BIN-676 | 11.9 | x | x* | " " |
| BIN-677 | 8.88 | x | x* | " " |
| BIN-678-1 | 6.37 | x | x* | " " |
| BIN-678-2 | 5.47 | x | x* | " " |
| BIN-695 | 6.06 | x | x* | " " |
| BIN-697 | 2.42 | ? | ? | " " |
| BIN-698 | 2.78 | ? | ? | " " |
| BIN-699 | 9.61 | x | x* | " " |
| BIN-700 | 0.55 | x | x* | " " |
| BIN-755 | 8.00 | x | x* | " " |
| BIN-572 | 16.7 | x | x* | Last Chance Fissure |
| BIN-605 | 11.4 | x | x* | " " " |
| BIN-614 | 17.1 | x | x* | " " " |
| BIN-619 | 7.15 | x | x* | " " " |
| BIN-620 | 11.8 | x | x* | " " " |
| BIN-622 | 9.15 | | x* | " " " |
| BIN-629 | 11.0 | x | x* | " " " |
| BIN-630 | 7.95 | | x* | " " " |
| BIN-635 | 9.30 | x | x* | " " " |
| BIN-621 | 11.8 | x | x* | D limestone |
| BIN-645 | 11.3 | x | x* | " " |
| BIN-651 | 16.1 | x | x* | " " |
| BIN-638 | 13.6 | x | x* | Niagara fissure |
| BIN-644 | 15.4 | x | x* | " " |
| BIN-674 | 7.25 | x | x* | " " |
| BIN-660 | 15.1 | x | x* | East fissure |
| BIN-662 | 4.20 | x | x* | " " |
| BIN-664 | 9.08 | x | x* | " " |
| BIN-666-1 | 13.2 | x | x* | " " |
| BIN-666-2 | 4.34 | x | x* | " " |
| BIN-656 | 16.5 | x | x | " " |
| BIN-701 | 0.58 | x | x* | Lark Vein |
| BIN-704 | 0.94 | x | x* | " " |
| BIN-706 | 0.26 | x | x* | " " |
| BIN-712 | 0.57 | x | x* | " " |
| BIN-723 | 0.88 | x | x* | Commercial limestone |
| BIN-732-1 | 1.69 | x | x* | " " |
| BIN-732-2 | 1.83 | x | x* | " " |
| BIN-734 | 2.08 | x | x* | Jordan limestone |

*Chalcopyrite as blebs in sphalerite

A considerably larger range of FeS content (1.96 to 17.9%) is evident in the remaining samples from the Central district. Most of these contain pyrite. Changes of temperature and sulfur pressure seem most likely to be responsible for these variations, but non-equilibrium is suggested as an additional possibility by the differences found within the polished sections of CEN-538A, 266, and 267. However, the iron content of these samples is low, and non-equilibrium may apply only to sphalerite deposited at relatively low temperatures (less than 300–400° C?) as suggested above.

An interesting trend of decreasing iron content away from the stocks is evident in the analyses. Figure 6 is an attempt to show this by plotting FeS content against distance to the Santa Rita or Hanover stock, whichever

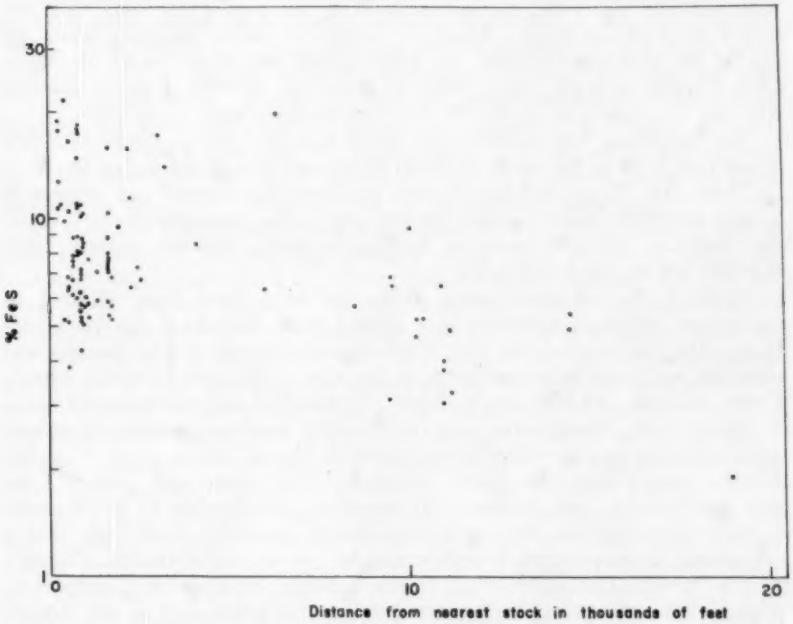


FIG. 6. Plot of FeS content of sphalerite against distance to nearest stock, showing decrease in FeS content with increasing distance from stock. Rank correlation coefficient is 0.44 compared to a critical value of 0.23 for 98% confidence.

is the closer. A significant rank correlation coefficient of 0.44 is found between these two variables, compared to the critical value of 0.23 for the 98% level of confidence (6). Although there is considerable variability around the trend, it is apparent that iron content is related in some way to the proximity of stocks. Temperature is an obvious cause for this relationship, but variation in chemical potential of sulfur is also a possible explanation. More information is needed to be certain of the cause of this difference.

## SPHALERITE FROM THE BINGHAM MINING DISTRICT

The sphalerite samples from Bingham were obtained from the U. S. and Lark sections of the U. S. Smelting, Refining and Mining Co. property. The production of these mines is dominantly lead with a smaller amount of zinc and minor amounts of silver, copper, and gold. The geology has been discussed by Hunt and Peacock (14), J. P. Hunt (15), and others. The ore represented by the analyzed sphalerite specimens occurs as replacements of thin limestone units in a sequence dominantly of quartzite, and as fissure fillings in intrusive rocks and sediments. Bedding in the U. S. section dips moderately to vertically, whereas in the Lark section bedding dips gently. Ore in several of the limestone units in the U. S. section has been followed down the dip of the bedding for over 3,000 feet. The ore within the bedding fissures generally occurs as shoots plunging down the dip of the structures toward the Utah Copper stock on which the large Utah Copper open-pit copper mine is centered. Shoots of pyrite coincide with the ore shoots in places, but in other places diverge from them.

The important ore minerals are galena and sphalerite. Pyrite is nearly ubiquitous, both in the separate shoots mentioned above and within the lead-zinc ore. In places, chalcopyrite and tetrahedrite-tennantite are present in notable amounts. Rarer minerals include pyrrhotite, marcasite, bornite, covellite, chalcocite, enargite, hematite, realgar, orpiment, cinnibar, stibnite, boulangerite, native silver and native gold.

Most of the limestone units within the mine have been silicated or marbleized, although relatively less altered black limestone can be found. Hunt (15) has studied the alteration minerals present in the fissures and adjoining wall rock in several parts of the mine. Minerals identified include garnet, diopside, wollastonite, feldspar, dioctahedral and trioctahedral micas of several types, dioctahedral and trioctahedral montmorillonoids of several types including montmorillonite and saponite, several kaolin group minerals, chlorite, vermiculite, talc, quartz, tridymite, cristobalite, opal, calcite, dolomite, rhodocrosite, and gypsum. Temperature classification of the deposits by either the sulfide or the non-sulfide mineral assemblages suggests that a large range of temperature is represented by the ore and alteration, although it seems possible that a few of the reported gangue minerals are supergene in origin. The widespread replacement and coarse grain size in the deeper levels seen by the writer suggest that mineralization processes were more intense at depth than in the upper levels, where much open-space filling but less replacement is evident.

The samples from the U. S. section analyzed in this study come from replacements of three limestone units and from five cross-fissures or fissure systems. Ore in the U. S. section is mostly medium to coarse-grained, the sphalerite and galena crystals ranging from a millimeter to several centimeters in size. Some ore is vuggy, with euhedral sulfide minerals projecting into the cavities. Banding parallel to fissures or bedding is defined by changes in mineralogy. In the Galena fissure thin color-banding occurs within zones of light-colored sphalerite, and is accompanied by thinly banded

mineralogical changes. In the other ores, the generally dark color of the sphalerite may vary, but rather than being banded, the color is concentrically zoned within single grains or shows irregular areas of dark and light sphalerite. In the suite of specimens from the Last Chance fissure, much of the sphalerite shows concentric color-zoning within single crystals, with light sphalerite in the center and dark sphalerite on the margins of grains. Two poorly developed examples of the reverse kind of zoning were noted, however. Color variations of sphalerite from other structures are of the irregular type where present, except one concentrically-zoned example from the B limestone footwall fissure. Sphalerite seen in polished sections frequently contains zonally arranged blebs of chalcopyrite.

Age relations of the sulfide minerals appear to be complicated. The sequence pyrite-sphalerite-tetrahedrite-galena fits most of the relations observed. The relations of chalcopyrite in various sections place it through the entire sequence of deposition from sphalerite through galena, and possibly also before sphalerite. Pyrite as veins and small masses shows a similar range in age from beginning to end of the sequence. Marcasite was deposited late in the sequence. The larger grains of chalcopyrite seem to have been deposited after sphalerite and before galena, but occurrences of large chalcopyrite grains are relatively rare in the sections examined.

In the Lark section east of the Utah Copper stock the ore comes largely from replacements and bedding fissures along three limestone units, the Jordan, Commercial and Lark limestones. Smaller amounts of ore are produced from cross fissures. The limestones dip gently northwest toward the Utah Copper stock. The ore being mined at present is mainly lead-zinc ore, but chalcopyrite is present in pyrite shoots and also in some galena and sphalerite ore. Much of the ore in the Lark section is associated with black limestone relatively less altered than limestones in the U. S. section. The sequence of deposition appears to be similar to that given for the U. S. section; at least such a sequence fits essentially all the relationships observed except that some tetrahedrite follows rather than precedes galena.

The sphalerite samples from Bingham can be conveniently divided into two groups. Samples from the B limestone, D limestone, Last Chance fissure, Niagara fissure, and East fissures form a group of dominantly high-iron sphalerites. Those from the Galena and Holden fissures, Highland, Jordan, and Commercial limestones, and Lark vein form a group of low-iron sphalerites. A similar grouping can be made on the basis of gallium content (22). FeS contents of 16.7% are reached in the high-iron group, with several values in the range of 14–16%, indicating minimum temperatures of 475 to 550° C. Such temperatures are consistent with the occurrence of garnet, diopside, wollastonite, and other pyrometasomatic minerals with some of the sphalerite. No pyrrhotite was found in the polished sections studied, although it has been reported from the district.

Figure 7 is a plot of the FeS content of sphalerites from the B limestone against elevation in the mine. A poorly developed trend toward lower iron content in the upper levels of the mine is evident. The sample at 1,600 feet may be a member of the low-iron, high gallium group more typical of the

Galena fissure and thus may not properly belong with the other specimens. A significant rank correlation coefficient of 0.69, compared to the critical value of 0.58 for 99% confidence, is found for this data (6). It seems likely that a temperature gradient is responsible for this trend, although perhaps in an indirect manner. The appearance of the ores suggests to the writer that replacement processes were more intense in the lower levels of the mine where the high-iron sphalerites occur, thus giving qualitative support to the hypothesis of a temperature gradient. The trend of individual ore shoots, downward toward the inferred position of the Utah Copper stock, is also consistent with this hypothesis.

The low-iron sphalerites from the Galena fissure, Lark vein, and elsewhere appear to be a good example of the effect of high sulfur pressure on the iron content of sphalerite. Abundant pyrite is present with most of these samples, indicating the presence of adequate iron, but most FeS contents are less than 2%, which is equivalent to a minimum temperature of less than 140° C. It seems unlikely that these sphalerites were deposited at such a low temperature, although they may have been deposited at lower temperatures than the high-iron group. Lack of equilibrium with the pyrite may add to the effect of high sulfur pressure in lowering the iron content.

### SPHALERITE FROM THE CLEVELAND MINE, WEST PINOS ALTOS DISTRICT, NEW MEXICO

At the Cleveland mine (now abandoned) in the West Pinos Altos district a few miles north of Silver City, most of the ore occurs as replacements of Pennsylvanian limestone (26). Typical ore runs 10% Zn, 1% Pb, 0.5% Cu, 1 oz. silver, and a trace of gold. The sphalerite is mainly coarse grained, crystals being commonly a quarter inch in diameter. Euhedral quartz is present in the ore specimens, and pyrrhotite in most of them. The contacts of ore with limestone are sharp, but a ½ to 2 inch layer of ankerite occurs between ore and limestone in many places.

The sphalerite almost invariably contains rounded blebs of chalcopyrite. In some sections, blebs and lamellae of pyrrhotite are also present, and a few larger grains of pyrrhotite were also observed. Lamellar marcasite is present in some sections, apparently having replaced pyrrhotite. Some chalcopyrite contains tiny star-shaped bodies of sphalerite.

The paragenesis is interpreted as follows, based on study of polished sections and hand specimens:

$$\text{Euhedral quartz} \text{—“Colloidal” pyrite—pyrrhotite—}$$

$$\text{sphalerite—chalcopyrite—}\begin{matrix}\text{galena}\\ \text{cosalite?}\end{matrix}\text{—pyrite}$$

The age of chalcopyrite and pyrrhotite is doubtful relative to sphalerite, but chalcopyrite definitely veins pyrrhotite. Marcasite replaces pyrrhotite and is veined by late pyrite.

Because of the consistently very high iron content of the sphalerite (Table 2a) plus the presence of pyrrhotite in most of the specimens, the temperatures indicated by the iron content are probably close to the actual temperature of deposition. The composition ranges from 18 to 24% FeS. Several of the analyzed sphalerite samples, mostly those of highest iron content, were observed to contain small amounts of pyrrhotite and chalcopyrite as impurities, largely as tiny blebs in the sphalerite. The maximum amount of pyrrhotite and chalcopyrite in the analyzed material is estimated to be 2 percent. A thin section shows that the sphalerite near these blebs is lighter in color than the main body of sphalerite, suggesting that all or part of the pyrrhotite and chalcopyrite blebs originated by exsolution from the nearby sphalerite. However, even if a correction for these inclusions is made by deducting them, the iron content falls in the range 18–22% FeS.

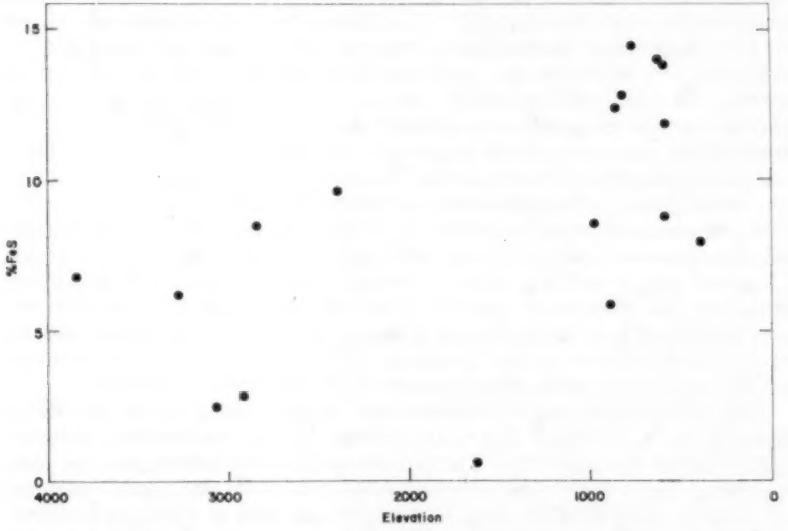


FIG. 7. Plot of FeS content of sphalerite samples from the B limestone against elevation in mine (relative to arbitrary datum). A decreased FeS content in the upper levels of the mine is suggested, and confirmed by a significant rank correlation coefficient of 0.69.

With no pressure correction a temperature of at least 650° C is indicated. Still higher temperatures are indicated if a pressure correction, equilibrium with non-stochiometric pyrrhotite, or exsolution of the blebs is allowed for.

It is interesting to compare the above temperature with the temperatures indicated by other minerals in the deposit. Chalcopyrite contains "stars" of sphalerite, suggesting exsolution and an origin at relatively high temperatures (9). Quartz occurs as prisms capped by various rhombohedrons, but no trapezohedral faces could be found, leaving unanswered the question of

whether it crystallized as high or low quartz. The chalcopyrite blebs, if originating by exsolution, also suggest high temperatures (5).

On the other hand, marcasite is common in some specimens. According to Allen, Crenshaw and Johnson (1), this mineral transforms to pyrite at a rapid rate above about 400° C. However, this mineral may have been deposited at a low-temperature stage of deposition considerably later than the sphalerite and pyrrhotite.

Ankerite-quartz intergrowths occur in the zinc ore, and calcite-quartz aggregates occur in the form of chert in limestone within half an inch of ore. No wollastonite or pyroxene was seen in hand specimens or in examination of half a dozen thin sections of ore and ore-wallrock contacts showing quartz-carbonate assemblages. With a $CO_2$ pressure equal to the rock pressure at a depth of 3,000 feet, wollastonite should form from calcite and quartz at a little under 600° C (12). For 6000 feet of overburden, the temperature would be about 625° C. The reaction of dolomite and quartz to form diopside or enstatite plus calcite would take place at about 375° C for 3,000 feet of overburden, and at a little over 400° C for 6,000 feet of overburden (27). Unfortunately, no data exist for predicting the temperature of reaction of iron-bearing dolomite (ankerite) plus quartz to form an iron-bearing pyroxene. Such a pyroxene (hedenbergite) did form at Hanover, and except in a few spots has not altered to carbonate and quartz or any other phases. The sphalerites of highest iron content at Hanover indicate temperatures of at least 550° C. Thus at Hanover it appears that hedenbergite was stable at about 600° and perhaps below, whereas at the Cleveland mine it was not formed at temperatures of at least 650° C. From these facts one is driven to conclude either that the $CO_2$ pressure at Hanover was very much lower than the rock pressure, despite the tremendous amounts of $CO_2$ being released as the limestones were replaced, or that at least one of the assemblages in question represents disequilibrium conditions.

An attempt was made to measure the lattice spacing of the pyrrhotite from the Cleveland mine and thus determine its composition and the prevailing sulfur pressure (2). The powder pattern obtained had fuzzy lines, suggesting that a zoned pyrrhotite was present, so no further work was done. However, a more detailed study of the pyrrhotite and of the deposit in general seems worthwhile in view of the high temperatures indicated by the sphalerite and the possibility of obtaining temperature data from other minerals.

## CONCLUSIONS

A number of recent papers have used the measured iron content of sphalerite to obtain temperatures of deposition. These temperatures have frequently been cited without discussion of the assumptions involved. An attempt has been made in this paper to point out the complications involved in the use of the iron content of sphalerite to measure temperature. The principal difficulties are believed to be lack of equilibrium, equilibrium with pyrite rather than pyrrhotite or equilibrium with non-stochiometric pyrrhotite

above 600° C, the pressure correction and possibly solid solution with chalcopyrite. Because almost all these situations lead to temperatures lower than the true temperature of deposition if the solvus for the FeS–ZnS system is used, it seems most reasonable to consider all sphalerites as giving minimum temperatures unless good evidence on all the factors mentioned above is available.

On this basis, deposition temperatures of at least 600° C were reached in the contact zone at Hanover, New Mexico, at least 300° C at the Groundhog mine in the same district, at least 650° C at the Cleveland mine in the nearby Pinos Altos district, and at least 550° C in the Bingham mining district, Utah. These temperatures are consistent with indications from most other minerals and mineral assemblages. However, late low temperature phases were apparently formed in all districts, and at the Cleveland mine higher than normal $CO_2$ pressures seem necessary to explain a lack of reaction between quartz and carbonates.

Differences of iron content within a single crystal and within several polished sections of sphalerite, even those containing pyrrhotite and deposited above 400° C, indicate that lack of equilibrium is not uncommon. Taking all sphalerites analyzed, regardless of the presence or absence of iron sulfides, a poorly developed but statistically significant lateral zoning of the iron content away from the igneous centers is found in the Central mining district. A similar poorly developed vertical zoning exists within one horizon at Bingham. In both cases temperature is probably involved as a cause, but may operate through control of the sulfur pressure rather than directly on the equilibrium iron content.

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BEAR CREEK MINING COMPANY,
SALT LAKE CITY, UTAH,
*June 3, 1961*

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# IRON ORE RESOURCES OF THE NORTH-EASTERN CONGO

R. A. WOODTLI

ABSTRACT

Iron ore deposits of good quality, related to belts of banded ironstones, occur in the northeastern part of the former Belgian Congo. Most of them have been known for more than half a century. Their geographical location made them economically unworkable and observations on their geological setting, their extent and composition were collected as a by-product of prospecting for primary gold deposits.

In the Upper Ituri Basin six main belts occur and are relatively well known. A first appraisal sets the reserves at about 900 million tons of ore above 45 percent iron.

Hematite of higher grade is widely distributed throughout the Upper Uelé Basin but a complete picture of their extent is still missing. As a first estimation of the reserves, the known data provide a total of 1,350 million tons of ore above 68 percent iron. The far eastern deposits, close to the border with Uganda, are the most readily accessible.

RÉSUMÉ

Depuis plus d'un demi-siècle on connaît des gisements de fer de bonne qualité dans le nord-est de l'ancien Congo belge. Leur éloignement de la mer ne permet pas encore d'envisager leur mise en valeur et ils n'ont guère fait l'objet de recherche. Les informations que nous possédons à leur sujet ont été recueillies à l'occasion de la prospection de gisements aurifères primaires.

Les gisements de fer les plus intéressants apparaissent sous forme de lentilles d'hématite plus ou moins pure enchâssées dans des itabirites. Ils sont relativement bien connus dans le Haut-Ituri où six zones principales d'itabirite ont été examinées. En première approximation, on peut fixer le montant des réserves à environ 900 millions de tonnes de minerai à plus de 45% de fer.

D'autres gisements, plus vastes et à meilleure teneur, sont abondamment représentés dans le Haut-Uelé où la région de Paulis a reçu le surnom de "Plateau de Fer." On est encore loin d'avoir une image complète de leur répartition. Les données connues permettent toutefois d'estimer les réserves à au moins 1.350 millions de tonnes de minerais à plus de 68% de fer.

Les gisements les plus à l'est, situés près de la frontière avec l'Uganda, sont les plus facilement exploitables.

As is well known iron ore formations are commonly associated with Precambrian rocks and the North-Eastern part of the Congo is no exception. Several conspicuous iron ore deposits occur in the territory drained by the upper course of the Ituri and Uele Rivers (Figs. 1, 2). The following is a short report on these deposits, based chiefly on the writer's observations.

Some of these ore bodies were discovered about half a century ago and

aroused interest since the first advent of the Europeans into this remote territory. Other discoveries were made in the course of prospecting for gold deposits, and in the course of geological mapping. Generally these observations were merely recorded without any detailed examinations of the ores. During the years the exploration of some gold deposits brought some valuable information on the iron formation in addition. It was obvious that their geographical location made them economically unworkable; their export was out of question and there were neither close coal deposits nor local market. The recent improvements of the direct reduction processes of iron ores, the existence of water power reserves in the territory and the need for industrial growth in the newly independent Congo will perhaps make these deposits more attractive in the near future.

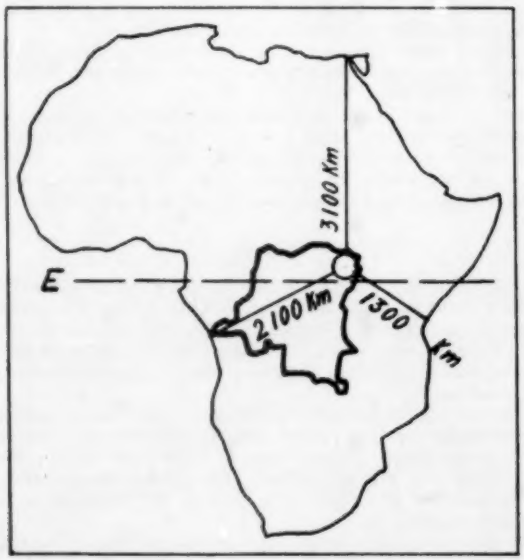
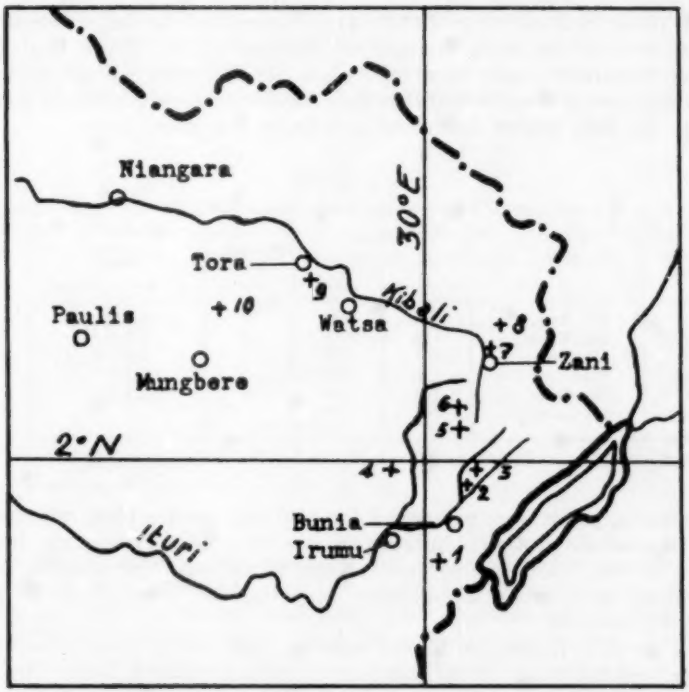



FIG. 1.

In the North-Eastern Congo the basement consists of granitoid rocks including large areas of highly metamorphosed rocks known as the "West Nile System" and of moderately metamorphosed formations thought to be Precambrian and called the "Kibali Group" or "Kibali System," the Kibali being the upper course of the Uele River. Ferruginous rocks locally called itabirites occur in the Kibalian and correspond to the banded ironstones that are widespread throughout Africa south of the Atlas Range. These formations occur in long and relatively narrow ridges that may be followed in some cases for tens of kilometers; many form a succession of lenses easy to recognize and to locate on aerial photographs. These belts are of some

value as "markers" in the mapping of structures. The elevation of the ridges above the surrounding country is extremely variable, ranging from 10 to 700 m. Their shape is unsymmetrical, with a steep slope on one side and the other one long and parallel to some resistant siliceous layer. They expose a typical profile of cuestas. Thick laterite crusts crown the top and extend around the foot of many of them. The itabirites are banded ferruginous quartzite and jaspilite, the iron content of which is generally too low




Key : 1 = Ami - 2 = Kuma - 3 = Go - 4 = Yedi
5 = Gobu - 6 = Lambiti - 7 = Kodo
8 = Maie - 9 = Gaima - 10 = Tinda.

FIG. 2.

to provide a workable iron ore. But certain portions are enriched by natural processes and may be classified in the category of high-grade ores. Weathering probably leached silica from the outcrops above the water table and enriched the ore in iron by oxidation of magnetite into hematite. The resulting lens-shaped bodies of massive and practically pure hematite are the

cleanest iron ore occurring in the territory. All these features seem to be common to Precambrian iron formations.

Though less abundant and poorer than those of the northern part of the district, the banded ironstones of the Upper Ituri Basin (or Kilo Region) are relatively well known, at least with regard to their distribution, outcrop aspect and composition of superficial rocks (Fig. 2). Besides the papers of Ancion and Cahen (1) and Cahen (3), which cover the broad field of Congolese iron resources, more detailed information may be gathered from the following publications: Duhoux (4), Woodtli (6, p. 20–28, map) and Woodtli (9, p. 71–77, Fig. 1, 13, 14). Figure 1 of the latter indicates the location of all important ferruginous outcrops of the Upper Ituri Basin. Other occurrences have been referred to W of Irumu but no recent and reliable report is available and even their location is in some instances doubtful.

In the Kilo Region the following belts may be mentioned:

TABLE 1

| No. on Fig. 2 | Name | Location | | Length km | Width km | Maximum elevation m |
|---|---|---|---|---|---|---|
| | | X | Y | | | |
| 1 | Mt. Ami | 228 | 709 | 12 | 0.2–3 | 1925 |
| 2 | Mt. Kuma | 251 | 759 | 12 | 0.1–0.4 | 1524 |
| 3 | Mt. Go | 261 | 782 | 12 | 0.2–0.5 | 1809 |
| 4 | Vedi | 196 | 783 | 2 | 0.1–0.2 | 1250 |
| 5 | Mt. Gobu | 243 | 815 | 10 | 0.1–0.5 | 1626 |
| 6 | Mt. Lambiti | 245 | 829 | ? | ? | 1497 |

N.B. Coordinates indicated are taken from Fig. 1 of Woodtli's paper. (9)

These deposits were prospected for gold and not for their iron content, but all available geological information was recorded. Prospecting methods consisted usually of detailed examinations of the outcrops, sampling across and along the outcrops, supplemented by pitting and trenching and in a few cases by tunneling.

In the Kilo Region the banded ironstone belts are in general concordant to the regional trend of the foliation of the surrounding rocks. In some instances they are disposed along an apparently abnormal plane between two formations, one of which commonly being granite. This is well marked in the region of Talolo (9, Figs. 6, 7, 13). Near Vieux Kilo, long and narrow ribbons of silicified ironstones are intersected by other similar bands with a displacement at the point of intersection (9, Fig. 14). These observations have aroused speculations about the mode of formation of the ironstones (7, 9).

Itabirites of higher grade are widely distributed throughout the Upper Uele Basin (or Moto region) but through lack of topographic and geological mapping, we have not yet a complete picture of their extent. In a few localities drilling has supplied detailed information on the rocks at depth. Much work has been done around Mt. Kodo near the small mining township of Zani,

in the far East (2, 6). P. Duhoux (4) indicates the following data for the Mt. Kodo deposit: length 1 km, width 0.4 km, elevation above sea level 1503 m. It consists mainly of pure hematite that extends some distance into Mt. Zani located to the South. However, the latter is composed chiefly of magnetite quartzite. In the same district, Mt. Maie with its top above 1600 m, offers another conspicuous outcrop of hematite surrounded by magnetite quartzite and other silicified ironstones.

In the region around Watsa and Tora several lines of hills correspond to itabirites and some of them enclose lenses of high-grade ores; the Gaima Mountain is such an example. North of Mungbere the Tina Range rises at least 500 m above the young Uele peneplain; its core is composed of a huge body of hematite. Farther to the West around Paulis extends the "Plateau de Fer" with many outcrops of ironstones of unknown economic value.

TABLE 2

UPPER ITURI BASIN: MORE THAN 45% FE ORES

| No. on Fig. 2 | Deposit | Length m | Width m | Height m | Density | Tonnage M.t. |
|---|---|---|---|---|---|---|
| 1 | Mt. Ami | 5.000 | 100 | 700 | 4.4 | 770 |
| 2 | Mt. Kuma | non workable | | | | |
| 3 | Mt. Go | 3.000 | 75 | 100 | 4.4 | 50 |
| 4 | Yedi | 2.000 | 100 | 100 | 4.4 | 25 |
| 5 | Mt. Gobu | 2.000 | 50 | 200 | 4.4 | 44 |
| 6 | Mt. Lambiti | data insufficient | | | | |
| | | | | | Total | 889 |

M.t. = millions of metric tons.

The drilling already mentioned has cast some light upon the mode of formation of these rocks (5, 8). Previously, P. Duhoux (4) has stated that the itabirites are the result of metasomatic replacement of a basement made either of Kibalian foliated rocks (micaschist, chlorite-schist, graphitic schist), or of ferrocarbonate bodies. The replaced rocks would supply the iron content of the itabirite. According to Duhoux the itabirite is a by product of the granitization of the substratum. Subsequent examination has shown a great extent of ankeritic and sideritic rocks below the water table; these latter are more or less strongly silicified and albitized. The problem of the origin of the carbonate rocks is not yet solved and several opinions have been expressed. However, it seems definitely demonstrated that the itabirites of the North-Eastern Congo come from the silicification and oxidation of ferrous carbonates. Secondary enrichment processes are responsible for the formation of high-grade ores.

Some banded ironstones belts are undoubtedly related to the emplacement of the granite; at least this is strongly suggested by their disposition on the margin of some granite bodies. In other cases they seem to be associated

with a pattern of radial faults (9). It must be added that only the first ones appear to bear some economic significance, as far as their iron content is concerned.

The published analyses (3) and unpublished data show that the ores generally contain no impurities, except silica. Manganese, nickel, chromium, titanum, phosphorus and sulfur are commonly very low or are missing. It must be noted that the ore composition even in the hematite lenses is known to change below the water table (8); carbonates, iron and arsenic sulfides are scattered through the rock and may be abundant enough to preclude any exploitation in the present technical conditions.

Though insufficient detailed work, particularly drilling, has been done to permit accurate estimate of the reserves in the deposits mentioned, a first appraisal based on the available data may provide a reasonably good approximation.

In the cases of Ami, Go, Gobu Mts, triangular longitudinal sections and rectangular cross sections have been considered in the calculations; in the

TABLE 3

UPPER UELE BASIN: MORE THAN 68% FE ORES

| No. on Fig. 2 | Deposit | Length m | Width m | Height m | Density | Tonnage M.t. |
|---|---|---|---|---|---|---|
| 7 | Mt. Kodo | 1.500 | 1.000 | 40 | 5 | 300 |
| 8 | Mt. Maie | 3.000 | 40 | 80 | 5 | 48 |
| 10 | Mt. Tina | 4.000 | 200 | 500 | 5 | 1.000 |
| | | | | | Total | 1.348 |

M.t. = millions of metric tons.

case of the Yedi deposit, a pyramidal shape has been admitted. Height is measured above the plain level.

In the Moto Region only pure hematite bodies are considered; in the cases of Kodo and Maie Mts a tabular shape is accepted in the calculations; for the Tina Range the shapes considered are a triangular longitudinal section and rectangular cross sections.

According to my observations the total of all known iron deposits in the North-Eastern Congo may be estimated at about 10 billion ton of ore above 45 percent iron.

The far Eastern deposits (7 and 8 of Fig. 2) are the more easily accessible at the present time. Some facilities are already available such as electrical power, good roads, trained man-power, cattle, and water. It seems unlikely that a steel industry will develop in this remote district, which lacks coal and limestone. But the Kodo high-grade ore may be extracted in open pit without removing barren rocks and with a minimum of preparatory work. These highly favorable conditions may make the ore attractive for export some day. In that case and in spite of the greater distance, transportation

to the Mediterranean sea using the Nile deserves attention. The second best deposit is the Tina ore located not too far from the railroad Aketi-Paulis-Mungbere.

GRANDVAUX,
SWITZERLAND.
*June 12, 1961*

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# URANIUM MIGRATION AND GEOCHEMISTRY OF URANIUM DEPOSITS IN SANDSTONE ABOVE, AT, AND BELOW THE WATER TABLE

## PART I.[1] CALCULATION OF APPARENT DATES OF URANIUM MIGRATION IN DEPOSITS ABOVE AND AT THE WATER TABLE[2]

JOHN N. ROSHOLT, JR.

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ABSTRACT

The migration of uranium may be studied by the distribution of the radioactive daughter products, which serve as natural tracers in the migration of uranium. The distribution of the daughter products is determined by radiochemical analyses of samples from ore deposits in sandstone, and the apparent minimum and maximum dates of uranium introduction or redistribution may be calculated from the $Pa^{231}/Th^{230}$ ratio. The primary assumption required is that the protactinium and thorium do not migrate in measurable quantities from the place where they were produced by the decay of the parent uranium isotopes. The upper limit of age determination is about 250,000 years, based on the half-lives of $Pa^{231}$ and $Th^{230}$. The difference in the half-lives of these isotopes is reflected in their differential rates of growth and decay corresponding to migrations of the parent uranium during the time range considered. The growth and decay patterns, analyzed mathematically, are used to determine the apparent date of uranium migration. Calculations based on analyses of samples from the Hulett Creek area, Wyoming, illustrate the results for typical sandstone ore deposits that are above and at the water table.

## INTRODUCTION

RADIOCHEMICAL analyses for radioactive equilibrium of several hundred samples from a wide variety of uranium deposits in sandstone, (3, 4) are proving useful in understanding the geochemical history of the deposits.

[1] Publication authorized by the Director, U. S. Geological Survey.
[2] Part I of two parts.

These analyses, when used to calculate the apparent age of the uranium in geologically controlled samples, allow interpretations as to the time and geochemical conditions of deposition of the uranium and subsequent leaching or enrichment of the deposits. The purpose of part I of this paper is to introduce methods of calculating the apparent maximum and minimum ages for deposits above and at the water table. These methods of calculation are

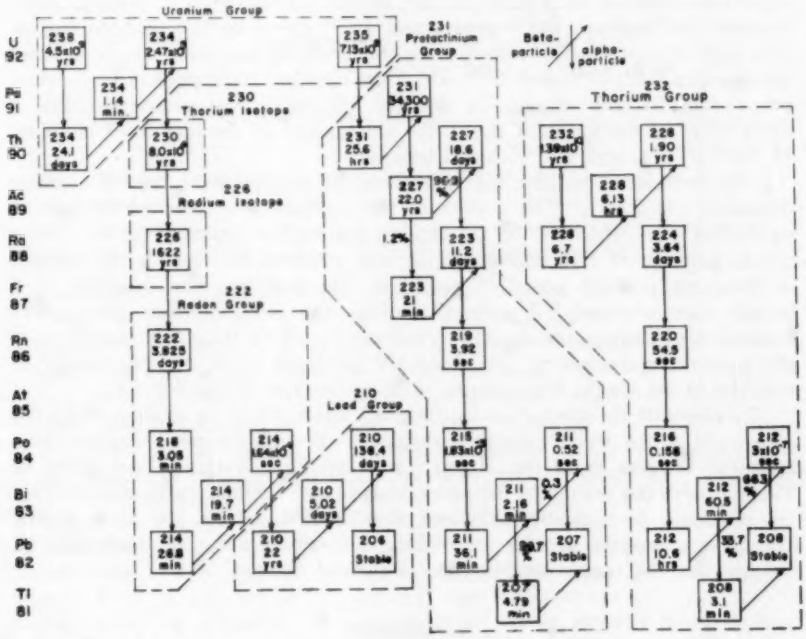


FIG. 1. Classification of natural radioisotopes into groups, showing atomic numbers, half-life periods, and modes of decay.

somewhat different than those previously published (5) for ground-water-saturated sandstone deposits. Part II of this paper is a discussion of the geochemical interpretations of the results of the calculations.

To make a detailed investigation of the state of radioactive equilibrium, several key decay produces must be measured and their relative abundances evaluated. Figure 1 shows the parent isotopes and decay or daughter products of the three principal naturally occurring radioactive decay series: the $U^{238}$, $U^{235}$, and $Th^{232}$ series. Equilibrium is attained in a radioactive series when all the daughter products decay at the same rate that they are produced from the parent isotope. Thus, at equilibrium each of these daughter products would be present in a constant proportion to its parent isotope.

The loss or gain, by geologic processes, of certain important isotopes during the more recent part of the existence of a mineral causes disequilibrium in the proportions of the parent isotope to its daughter products. Even though a series is not in complete equilibrium, many of the immediate short-lived daughter products will be in equilibrium with their long-lived parents. Where significant disruption of equilibrium occurs, the natural radioisotopes can be separated into the following major isotopes and groups of established equilibrium shown in Figure 1: the uranium group, $Th^{230}$ isotope, $Ra^{226}$ isotope, $Rn^{222}$ group, $Pb^{210}$ group, and $Pa^{231}$ group in the uranium series. The $U^{235}$ content is determined from the uranium analyses, which includes the isotopes $U^{238}$, $U^{235}$, and $U^{234}$, which are assumed to be present in constant proportions to each other. To describe adequately the long-term state of equilibrium of the uranium series, the abundances of the long-lived isotopes, U, $Pa^{231}$, $Th^{230}$, and $Ra^{226}$, should be known.

The methods of making the radiochemical analyses have been previously described (1, 2, 6). The results of the analyses are reported in percent equivalent (2), which is the amount, in percent, or primary parent, under the assumption of radioactive equilibrium, required to support the amount of daughter product actually present in the sample. For example, if a sample that contained 0.2 percent uranium was in equilibrium, the percent equivalent of each of the daughter products would be 0.2. The variation of the percent equivalent of the daughter products from the percentage of uranium in the sample is a measure of the radioactive disequilibrium.

To illustrate the method of apparent age calculations for uranium deposits above and at the water table, the results of the radiochemical analyses of a series of samples from the Hulett Creek area in Wyoming are given in Table 1, with the results of the age calculations. Included with the analyses for uranium, protactinium-231, and thorium-230, which are used in the calculations, are the results for radium-226, radon-222, and lead-210. A definition of the terms used in the calculations follows.

*Nomenclature*

$\lambda$ = decay constant of $Th^{230} = 0.866 \times 10^{-5}$ per year.
$\beta$ = decay constant of $Pa^{231} = 2.02 \times 10^{-5}$ per year.
U = actual uranium content of sample in percent.
$Pa^{231}$ = actual protactinium-231 content of sample in percent equivalent.
$Th^{230}$ = actual thorium-230 content of sample in percent equivalent.
$U_1$ = single generation uranium content.
$U_0$ = first generation uranium content of samples with second generation uranium.
$U_2$ = second generation uranium content.
$Th_1^{230}$, $Pa_1^{231}$ = present isotope abundance created from first generation uranium.
$Th_2^{230}$, $Pa_2^{231}$ = present isotope abundance created from second generation uranium.
$t_{Pa}$ = theoretical time required for growth of $Pa^{231}$ in $Pa^{231}/U$ ratio.
$t_{Th}$ = theoretical time required for growth of $Th^{230}$ in $Th^{230}/U$ ratio.
$t_2$ = maximum date of introduction of second generation uranium.
$t_1$ = date of leaching of first generation uranium.
$\Delta t$ = correction, in years, for residual uranium content.

The calculations of apparent age, and the interpretations regarding the migration of uranium, up to a maximum time of 250,000 years, are made with the use of protactinium-231 and thorium-230 abundances in combination with the uranium content of the sample. Protactinium-231 has a half life of 34,000 years and thorium-230 has a half life of 80,000 years.

TABLE 1. CHEMICAL AND RADIOCHEMICAL ANALYSES AND AGE CALCULATIONS OF SAMPLES FROM HULETT CREEK AREA, WYOMING

| | | PERCENT | | PERCENT EQUIVALENT | | | | | CALCULATED RESULTS FOR $U_0$ OR $U_1$ 1/ | | | | CALCULATED RESULTS FOR $U_2$ 1/ | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | | | | | YEARS | | WEIGHT % | | YEARS | | WEIGHT % | |
| LABORATORY NUMBER | FIELD NUMBER | eU | U | $Pa^{231}$ | $Th^{230}$ | $Ra^{226}$ | $Rn^{222}$ | $Pb^{210}$ | minimum | maximum | U min | U max | min | max | U min | U max |
| 252409 | 79CR56 | 1.4 | 2.11 | 2.23 | 2.16 | 1.90 | 0.97 | 1.62 | 220,000 | >250,000 | 4.0 | 2.3 | | | | |
| 252410 | 80CR56 | .19 | .34 | .41 | .20 | .16 | .13 | .13 | 15,000 | 25,000 | 2.8 | 1.0 | | | | |
| 252411 | 81CR56 | .22 | .40 | .26 | .22 | .21 | .11 | .15 | 81,000 | >250,000 | .21 | .18 | 0 | 22,000 | .40 | .22 |
| 252412 | 82CR56 | .20 | .38 | .25 | .16 | .19 | .11 | .14 | 54,000 (CONCORDANT DATE) | 62,000 (CONCORDANT DATE) | .38 | .38 | | | | |
| 252418 | 112CR56 | 1.7 | 3.06 | 3.33 | 3.27 | 2.84 | 1.5 | .73 | 200,000 | >250,000 | 6.1 | 3.4 | | | | |
| 252419 | 113CR56 | .41 | .69 | .84 | .53 | .40 | .29 | .17 | 55,000 | 84,000 | 2.1 | 1.1 | | | | |
| 252423 | 86CR56 | .21 | .25 | .15 | .23 | .18 | .13 | .17 | 160,000 | >250,000 | .92 | >.60 | 0 | 45,000 | .25 | .25 |
| 252424 | 87CR56 | .51 | .13 | .89 | .74 | .93 | .65 | .84 | 81,000 | 160,000 | 3.5 | 1.0 | | | | |
| 252425 | 88CR56 | .10 | .043 | .12 | .19 | .15 | .14 | .15 | >250,000 | >250,000 | .28 | .28 | | | | |
| 252426 | 89CR56 | .029 | .010 | .041 | .035 | .032 | .04 | .06 | 87,000 | 210,000 | .15 | .04 | | | | |
| 252427 | 90CR56 | .28 | .45 | .25 | .41 | .27 | .17 | .17 | 165,000 | >250,000 | 1.6 | >.29 | 0 | 40,000 | .45 | .45 |
| 252428 | 91CR56 | .12 | .15 | .085 | .050 | .028 | .03 | .03 | 21,000 (CONCORDANT DATE) | 25,000 (CONCORDANT DATE) | .15 | .15 | | | | |
| 252429 | 92CR56 | .15 | .16 | .19 | .22 | .18 | .14 | .17 | >250,000 | >250,000 | .26 | .26 | | | | |
| 252430 | 93CR56 | .23 | .034 | .97 | .18 | .38 | .32 | .37 | <2,000 | <2,000 | -- | -- | | | | |
| 252431 | 94CR56 | 2.1 | 1.56 | 1.64 | 1.94 | 2.42 | 2.33 | 2.40 | >250,000 | >250,000 | 2.8 | 2.8 | | | | |
| 252432 | 97CR56 | .10 | .016 | .31 | .22 | .21 | .21 | .19 | 61,000 | 123,000 | 1.2 | .34 | | | | |
| 252433 | 98CR56 | .17 | .18 | .12 | .21 | .18 | .15 | .17 | 175,000 | >250,000 | .86 | >.55 | 0 | 55,000 | .18 | .18 |
| 252434 | 99CR56 | .12 | .17 | .088 | .092 | .070 | .07 | .058 | 94,000 | >250,000 | .39 | .08 | 0 | 6,000 | .12 | .09 |

1/ $U_0$, SINGLE GENERATION URANIUM CONTENT; $U_1$, FIRST GENERATION URANIUM CONTENT WITH SECOND GENERATION URANIUM; $U_2$, SECOND GENERATION URANIUM CONTENT.

ANALYSTS: eU, D. L. SCHAFER; U, H.J. PENNELLY; RADIOCHEMICAL, J.N. ROSHOLT, JR.

## METHODS OF AGE CALCULATION

To make the calculations, and to make further interpretations, two fundamental assumptions must be made; (1) that the $Pa^{231}$ and $Th^{230}$ isotopes do not migrate in measurable amounts away from the place where they were created by the radioactivity decay of their parent uranium isotopes, and (2) that the uranium leached from the original amount present in the rock sampled was not leached out prior to a time that would be one-half of the total time the uranium had resided in that particular specimen of rock. The second assumption is necessary to establish a reasonable boundary condition in order to estimate minimum dates of uranium introduction.

The relationship of equivalent protactinium-231 to equivalent thorium-230 in any sample may be described as the protactinium-231 being equal to, greater than, or less than the thorium-230. When the uranium content of the sample is considered, nine possible abundance relationships exist, as shown on Figure 2. The nine abundance relationships may be divided, based on relative decay rates of the uranium and its daughter products, into two groups; (1) indicating that there need be only one generation of uranium deposition, with the possibility of some subsequent leaching of uranium, to produce the abundance relationships found, and (2) indicating that the

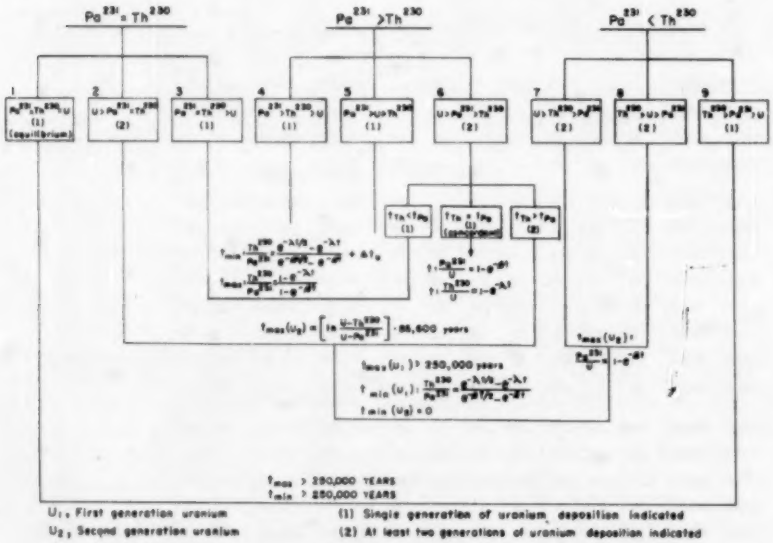


FIG. 2. Classes and minimum-maximum date relationships for U, $Pa^{231}$, $Th^{230}$ combinations.

abundance relationships must have been the result of at least two generations of uranium deposition. Of the nine abundance relationships (Fig. 2), cases 1, 3, 4, 5, and 9 belong to group 1 and all, with the exception of case 1—the equilibrium case—require some subsequent leaching of the uranium. Cases 2, 7, and 8 require at least a second generation of uranium deposition and belong to group 2. Case 6, where $U > Pa^{231} > Th^{230}$, requires an additional test to determine if a second generation of uranium is required.

The results of sample analyses where $U > Pa^{231} > Th^{230}$ (case 6, Fig. 2) are further tested by determining the relationship of the theoretical time required for the growth of $Pa^{231}$ in the $Pa^{231}/U$ ratio ($t_{Pa}$) to the theoretical time for the growth of $Th^{230}$ in the $Th^{230}/U$ ratio ($t_{Th}$). A plot of the

variation of the ratio of $Pa^{231}/U$ and $Th^{230}/U$ with time is shown in Figure 3 where:

$$Pa^{231}/U = 1 - e^{-\beta t_{Pa}} \tag{1}$$

and

$$Th^{230}/U = 1 - e^{-\lambda t_{Th}} \tag{2}$$

The results of this test may be divided into three subcases, which are shown on Figure 2 below case 6. When $t_{Pa} \cong t_{Th}$, within 20 percent, a concordant data is indicated requiring only a single generation of uranium deposition.

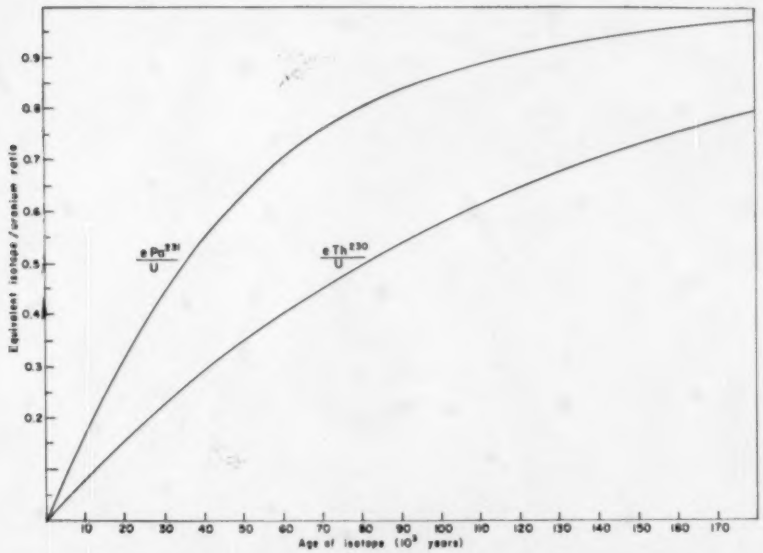


FIG. 3. Time relationships of $Pa^{231}$ to U and $Th^{230}$ to U.

When $t_{Pa} > t_{Th}$, a single generation of uranium is required with some subsequent uranium leaching. When $t_{Pa} < t_{Th}$, a second generation of uranium deposition is required.

The deposits of the Hulett Creek area, Wyoming, were chosen to illustrate the methods of age calculation because samples from these deposits represent most of the cases of abundance relationships (Fig. 2), except for case 1, the equilibrium case. Case 2 can be considered to be represented by sample 99 and case 3 by samples 79 and 112 as the difference between the percent equivalent $Pa^{231}$ and $Th^{230}$ is less than 5 percent (Table 1). Case 4 is represented by samples 87, 89, 93, and 97, and case 5 by samples 80 and 113. of the subcases of case 6 (Fig. 2), the concordant dates where $t_{Pa} = t_{Th}$ is represented by samples 82 and 91, and where $t_{Pa} < t_{Th}$ by sample 81. Case 7

is represented by samples 86 and 90, case 8 by sample 98, and case 9 by samples 88, 92, and 94.

Apparent maximum and minimum dates for the introduction of the uranium are calculated as it is probable that the uranium was introduced over a period of time rather than at a specific time. The agreement between the apparent maximum and minimum dates and the actual maximum and minimum dates for the introduction of the uranium will be dependent upon the validity of the two assumptions on which the calculations are based for

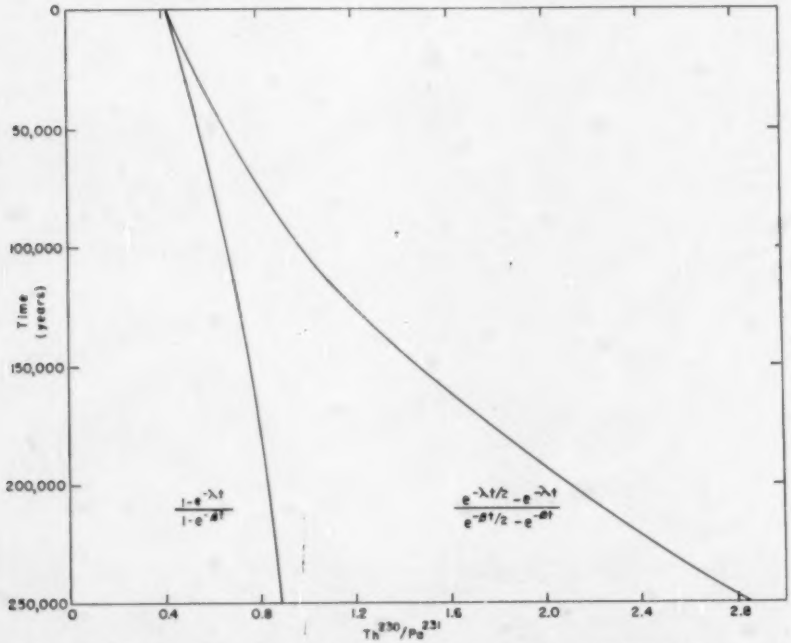


FIG. 4. Minimum-maximum time relationships based on $Th^{230}/Pa^{231}$ ratio.

that particular sample. That is, whether there has been any migration of the $Pa^{231}$ and $Th^{230}$ isotopes, and the amount that may have migrated, since the deposition of the uranium, and whether any uranium has been leached, and the amount leached, prior to a time equal to one-half the time the uranium has resided in the rock sampled. No additional limits of error are included with the result since the use of minimum and maximum dates define the possible range of error.

*Maximum and Minimum Dates for Single Generation Uranium.* The maximum date for the introduction of the uranium in those cases where only a single generation of uranium is required, would be dependent only on the

ratio of $Th^{230}/Pa^{231}$. This may be determined from a plot of the following function against time, shown on Figure 4:

$$\frac{Th^{230}}{Pa^{231}} = \frac{1 - e^{\lambda t}}{1 - e^{\beta t}} \tag{3}$$

For example, for sample 89 (Table 1), it can be determined from Figure 4 that the apparent maximum age of this sample is about 210,000 years.

The apparent minimum date can be obtained, if all the uranium was leached out when the deposit was half its minimum age, with a plot of the following function against time, which is shown on Figure 4:

$$\frac{Th^{230}}{Pa^{231}} = \frac{(1 - e^{-\lambda t/2})(e^{-\lambda t/2})}{1 - e^{-\beta t/2}(e^{-\beta t/2})} = \frac{e^{-\lambda t/2} - e^{\lambda t}}{e^{-\beta t/2} - e^{\beta t}} \tag{4}$$

Using sample 89 as an example again and Figure 4, the apparent minimum age is about 81,000 years.

All of the uranium leached samples, however, have some uranium left, which would increase the value of the apparent minimum date. The true apparent minimum date when the uranium content of the sample is taken into consideration can be derived as follows:

$$Th^{230} = U_0(1 - e^{-\lambda t/2})(e^{-\lambda t/2}) + U(1 - e^{-\lambda t/2}) \tag{5}$$

$$Pa^{231} = U_0(1 - e^{-\beta t/2})(e^{-\beta t/2}) + U(1 - e^{-\beta t/2}) \tag{6}$$

Thus

$$U_0 = \frac{Th^{230} - U(1 - e^{-\lambda t/2})}{e^{-\lambda t/2} - e^{-\lambda t}} = \frac{Pa^{231} - U(1 - e^{-\beta t/2})}{e^{-\beta t/2} - e^{-\beta t}} \tag{7}$$

and

$$\frac{e^{-\lambda t/2} - e^{-\lambda t}}{e^{-\beta t/2} - e^{-\beta t}} = \frac{Th^{230} - U(1 - e^{-\lambda t/2})}{Pa^{231} - U(1 - e^{-\beta t/2})} \tag{8}$$

The solution of equation (8) for sample 89 gives a value of about 6,000 years, which added to the value obtained from Figure 4 gives a true apparent minimum age of 87,000 years.

An approximate date correction for the residual uranium for the apparent minimum age can be more easily obtained, than by the solution of equation (8), by using the nomograph, Figure 5. The intersection of a line between the $Th^{230}/Pa^{231}$ ratio and the ratio of the residual U to the average $Pa^{231}$, $Th^{230}$ value with the empirical $\Delta t$ plot gives the approximate date correction. For sample 89 the use of Figure 5 gives a date correction of about 5,000 years. This added to the result from equation (4), or Figure 4, gives an apparent minimum age of 85,000 years as compared to the calculated apparent minimum age of 87,000 years.

The original uranium content of the sample that would be required at

the apparent age can be calculated from the following function:

$$U_0 = \frac{Th^{230}}{1 - e^{-\lambda t}} \tag{9}$$

The original uranium content that would be required at the apparent minimum date can be calculated from equation (7). In Table 1, the percent uranium at the maximum apparent age is listed as $t_{\text{maximum U}}$ and the percent of uranium at the apparent minimum age as $t_{\text{minimum U}}$. For sample 89, $t_{\text{maximum U}}$ from equation (9) is 0.04 percent, and $t_{\text{minimum U}}$ from equation (7) is 0.15 percent.

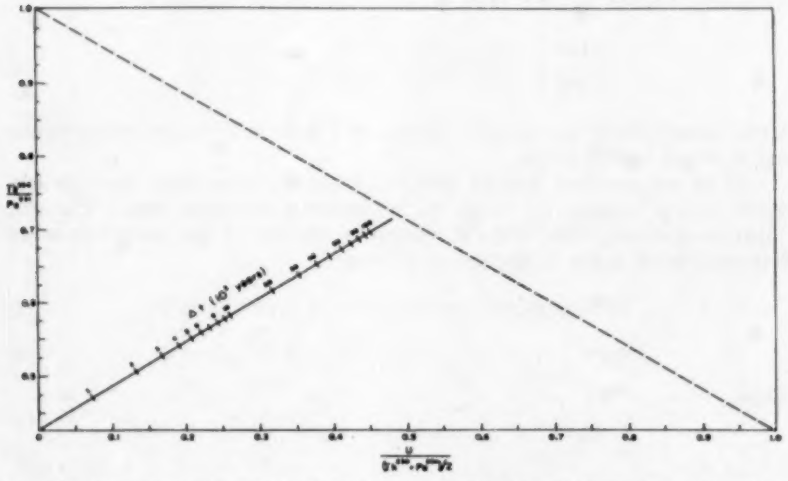

FIG. 5. Nomograph for residual uranium correction for minimum date estimation.

These figures of the uranium concentration at a particular time, particularly at the apparent minimum age, are important in understanding the geochemical history of a deposit, which will be discussed in part II of this paper. But, it will be of interest here to point out that the concentration of uranium in the deposit represented by sample 89 has decreased from 0.15 percent 87,000 ago to 0.010 percent at present, if the minimum age is assumed to be the correct age.

*Maximum and Minimum Dates for Second Generation Uranium.* A second generation of uranium is indicated for those deposits in which $U > Pa^{231}$ and $Pa^{231} \leqq Th^{230}$ (Fig. 2). The results of the calculations of the apparent maximum and minimum dates of introduction of the first and second generations of uranium, and the amounts of uranium deposited, are not as definitive as with a single generation of uranium, but some minimum and maximum can be determined.

The apparent minimum date for the introduction of the first generation uranium is determined by the same method (equation 4 and Fig. 4) as for

single generation uranium. The figure obtained will be somewhat lower than the actual minimum date for the introduction of the first generation uranium because, by this method of calculation, we assume that the $Th^{230}/Pa^{231}$ ratio is the result of a single generation of uranium deposition, and that all of the first generation uranium was leached out when the deposit was half its minimum age. A correction for the residual first generation uranium, as is done for the single generation uranium (equation 8 or Fig. 5), cannot be made because of the addition of the second generation uranium. As used, the $Th^{230}/Pa^{231}$ ratio defines the minimum date for the introduction of the second generation uranium would be 0 because theoretically no $Th^{230}$ or $Pa^{231}$ have been produced by the second generation uranium.

As an example, for sample 86 (Table 1) using Figure 4, we obtain 160,000 years as the minimum date for the introduction of the first generation uranium, and the minimum date for the introduction of the second generation uranium is 0.

For those deposits where two generations of uranium are indicated, the apparent date of the introduction of the first generation of uranium becomes older as the date of the second generation of uranium becomes older until the first generation of uranium has reached equilibrium. This is the result of the decay rate of $Th^{230}$ being less than half that of $Pa^{231}$, and the relative proportion of $Th^{230}$ to $Pa^{231}$ attributed to the first generation uranium increases with the increase of daughter products with increase of age of the second generation of uranium. It may be seen from Figure 4, where $(1-e^{\lambda t})/(1-e^{\beta t})$ is plotted against time, that an increase in the ratio of $Th^{230}/Pa^{231}$ will increase the age. When it is assumed that the first generation of uranium has reached equilibrium, and the date of introduction therefore greater than 250,000 years, then the apparent maximum date of introduction of the second generation of uranium can be derived for two different conditions.

The first condition will exist when $U > Pa^{231} > Th^{230}$ and when $t_{Th} > t_{Pa}$ (Fig. 2) then:

$$U_0 = Pa_1^{231} = Th_1^{230} \tag{10}$$

$$Th_2^{230} = U_2(1 - e^{-\lambda t_2}) \tag{11}$$

$$Pa_2^{231} = U_2(1 - e^{\beta t_2}) \tag{12}$$

$$U = U_1 + U_2 \tag{13}$$

$$Th^{230} = Th_1^{230} + Th_2^{230} = Pa_1^{231} + (U - Pa_1^{231})(1 - e^{-\lambda t_2}) \tag{14}$$

$$Pa^{231} = Pa_1^{231} + Pa_2^{231} = Pa_1^{231} + (U - Pa_1^{231})(1 - e^{-\beta t_2}) \tag{15}$$

$$Th^{230}e^{\lambda t_2} - U(e^{\lambda t_2} - 1) = Pa^{231}e^{\beta t_2} - U(e^{\beta t_2} - 1) \tag{16}$$

$$\frac{e^{\beta t_2}}{e^{\lambda t_2}} = \frac{Th^{230} - U}{Pa^{231} - U} \tag{17}$$

$$t_2 = \frac{1}{\beta - \lambda} \ln \frac{Th^{230} - U}{Pa^{231} - U} = 86{,}600 \left[ \ln \frac{U - Th^{230}}{U - Pa^{231}} \right] \text{ years} \tag{18}$$

Sample 81 (Table 1) fulfills these conditions, and using equation (18) the apparent maximum age of the secondary uranium is about 22,000 years. If the assumption that the first generation uranium is in equilibrium is not valid, the maximum date for the introduction of the second generation of uranium, as calculated from equation (18), will be too old in proportion to the degree of disequilibrium.

The second condition will exist when $U > Th^{230} > Pa^{231}$ and $Th^{230} > U > Pa^{231}$ (Fig. 2). The apparent maximum date for the second generation of uranium can be calculated from equation (1) and Figure 3 where:

$$1 - e^{-\beta t_2} = \frac{Pa^{231}}{U} \tag{19}$$

if $U_2$ is assumed to be equal to U. Samples 86 and 98 (Table 1) fulfill these conditions. Using Figure 3, the apparent maximum date of the introduction of the second generation of uranium in sample 86 is 45,000 years and for sample 98 it is 55,000 years.

If U is not equal to $U_2$, as assumed, and there is some $U_1$ in the sample, the maximum date as calculated is too old in proportion to the amount of $U_1$ present.

The amount of first and second generation uranium can be determined for the conditions used in calculating the apparent minimum and maximum dates for the introduction of the uranium. For the first generation uranium, no finite value can be calculated without making assumptions as to the time when all the $U_1$ was leached out and the degree of radioactive equilibrium reached prior to the leaching. The minimum value for $U_1$, however, would be represented by the function:

$$Th^{230} = U(1 - e^{\lambda t_2}) + U_1(e^{-\lambda t_1}) \tag{20}$$

and since $t_1$ must be such that $e^{-\beta t_1} < 0.04$, then $t_1 \geq 160{,}000$ years, and $U_1$ will be:

$$U_1 = \frac{Th^{230} - U(1 - e^{-\lambda t_2})}{0.25} \tag{21}$$

When $U > Pa^{231} > Th^{230}$ and $t_{Th} < t_{Pa}$ by using equations (10, 11, 13, and 14) it follows that:

$$U_2 = \frac{U - Th^{230}}{e^{-\lambda t_2}} \tag{22}$$

and

$$U_1 = U - U_2 \tag{23}$$

When $U > Th^{230} > Pa^{231}$ and $Th^{230} > U > Pa^{231}$ it is assumed that:

$$U_2 = U \tag{24}$$

The methods of calculating the maximum and minimum age of the introduction of the uranium, and the amount of uranium introduced, given on

the preceding pages, were used to calculate the values given in Table 1. To evaluate these results, and to show their significance in the geochemistry of uranium deposits, the geology and geochemistry of the deposits from which the samples were taken are described in part II of this paper.

U. S. GEOLOGICAL SURVEY,
DENVER, COLORADO,
*May 29, 1961*

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Vol. 56, 1961, pp. 1404–1420

# URANIUM MIGRATION AND GEOCHEMISTRY OF URANIUM DEPOSITS IN SANDSTONE ABOVE, AT, AND BELOW THE WATER TABLE
## PART II. RELATIONSHIP OF URANIUM MIGRATION DATES, GEOLOGY, AND CHEMISTRY OF THE URANIUM DEPOSITS[1]

CHARLES S. ROBINSON AND JOHN N. ROSHOLT, JR.

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ABSTRACT

The time of uranium migration in deposits in sandstone can be determined by correlating apparent age calculations, based on radiochemical analyses, with the geology of a particular deposit. Data were obtained from uranium ore samples representing deposits above the water table, deposits just above and below perched water tables, and deposits at least 250 feet below the water table in the Hulett Creek area, Wyoming. The first uranium deposition occurred more than 250,000 years ago for the deposits now at or above the water table. Approximately 60,000 to 80,000 years ago these deposits were oxidized, leached, and locally enriched. Accumulation of uranium in the deposits below the water table probably did not start before 180,000 years ago and has continued to the present.

## INTRODUCTION

THE apparent ages of uranium introduction in deposits in sandstone, based on radiochemical analyses, as previously published (9) and in part I of this paper, range from 0 to more than 250,000 years. An examination of the results for a series of samples, such as given in Table 1, part I, would seem to indicate

[1] Publication authorized by the Director, U. S. Geological Survey.

TABLE 1. AGE CALCULATIONS BASED ON RADIOCHEMICAL ANALYSES OF SAMPLES FROM THE HULETT CREEK AREA, WYOMING

| LABORATORY NUMBER | FIELD NUMBER | PERCENT U | CALCULATED RESULTS FOR $U_0$ OR $U_1$ 1/ | | | | CALCULATED RESULTS FOR $U_2$ 2/ | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | YEARS | | WEIGHT % | | YEARS | | WEIGHT % | |
| | | | $t_{minimum}$ | $t_{maximum}$ | U $t_{min}$ | U $t_{max}$ | $t_{min}$ | $t_{max}$ | U $t_{min}$ | U $t_{max}$ |
| | | | DEPOSITS ABOVE THE WATER TABLE (CARNOTITE-TYUYAMUNITE DEPOSITS) | | | | | | | |
| 252409 | 79CR56 | 2.11 | 220,000 | >250,000 | 4.0 | 2.3 | | | | |
| 252410 | 80CR56 | .34 | 15,000 | 25,000 | 2.8 | 1.0 | | | | |
| 252411 | 81CR56 | .40 | 81,000 | >250,000 | .21 | .18 | 0 | 22,000 | 0.40 | 0.22 |
| 252412 | 82CR56 | .38 | 54,000 | 62,000 | .38 | .38 | | | | |
| 252418 | 112CR56 | 3.06 | 200,000 | >250,000 | 6.1 | 3.4 | | | | |
| 252419 | 113CR56 | .69 | 55,000 | 84,000 | 2.1 | 1.1 | | | | |
| | | | DEPOSITS AT THE WATER TABLE (PYRITE OXIDIZED) | | | | | | | |
| 252424 | 87CR56 | .13 | 81,000 | 160,000 | 3.5 | 1.0 | | | | |
| 252426 | 89CR56 | .010 | 87,000 | 210,000 | .15 | .04 | | | | |
| 252430 | 93CR56 | .034 | <2,000 | <2,000 | -- | -- | | | | |
| 252432 | 97CR56 | .016 | 61,000 | 123,000 | 1.2 | .34 | | | | |
| | | | DEPOSITS AT THE WATER TABLE (PYRITE UNOXIDIZED) | | | | | | | |
| 252423 | 86CR56 | .25 | 160,000 | >250,000 | .92 | >.60 | 0 | 45,000 | .25 | .25 |
| 252425 | 88CR56 | .043 | >250,000 | >250,000 | .28 | .28 | | | | |
| 252427 | 90CR56 | .45 | 165,000 | >250,000 | 1.6 | >.29 | 0 | 40,000 | .45 | .45 |
| 252428 | 91CR56 | .25 | 21,000 | 25,000 | .15 | .15 | | | | |
| 252429 | 92CR56 | .16 | >250,000 | >250,000 | .26 | .26 | | | | |
| 252431 | 94CR56 | 1.56 | >250,000 | >250,000 | 2.8 | 2.8 | | | | |
| 252433 | 98CR56 | .18 | 175,000 | >250,000 | .86 | >.55 | 0 | 55,000 | .18 | .18 |
| 252434 | 99CR56 | .12 | 94,000 | >250,000 | .39 | .08 | 0 | 6,000 | .12 | .09 |
| | | | DEPOSITS BELOW WATER TABLE (HAUBER MINE) | | | | | | | |
| 269010 | H1 | .048 | 23,000 | 52,000 | | | | | | |
| 269011 | H2 | .33 | 46,000 | 130,000 | | | | | | |
| 269012 | H3 | .56 | 41,000 | 85,000 | | | | | | |
| 269013 | H4 | 1.30 | 57,000 | 180,000 | | | | | | |
| | WILL PULP LOT NUMBERS | | | | | | | | | |
| 271915 | 188 | .17 | 33,000 | 80,000 | | | | | | |
| 271916 | 189 | .19 | 41,000 | 85,000 | | | | | | |
| 271917 | 190-191 | .16 | 28,000 | 62,000 | | | | | | |
| 271918 | 192-193 | .17 | 45,000 | 130,000 | | | | | | |

1/ $U_0$, SINGLE GENERATION URANIUM CONTENT; $U_1$, FIRST GENERATION URANIUM CONTENT WITH SECOND GENERATION URANIUM; $U_2$, SECOND GENERATION URANIUM CONTENT.

ANALYSTS: U, H.J. FENNELLY, H.H. LIPP, AND D.L. FERGUSON

that there was little significance to the results, as the range is throughout the limits of the methods of calculations. When the geology of the samples is considered, however, it is possible to evaluate the results. The purpose of this part of the paper is to relate the apparent age results to the date of uranium

migration; to relate these dates to the geology of the deposits; and to show that the results are pertinent to the origin and migration of uranium deposits in sandstone. This study also indicates that it is possible to use radiochemical analyses and apparent age calculations in the search for uranium deposits. The samples considered are from the Hulett Creek area of northeastern Wyoming.

The data presented on the deposits are far from complete, and the interpretations, therefore, preliminary. The authors present the data and their interpretations at this time in the hope that it may stimulate others in pursuing what promises to be a profitable line of research. No effort is made to extrapolate the data to other uranium deposits in sandstone in the Black

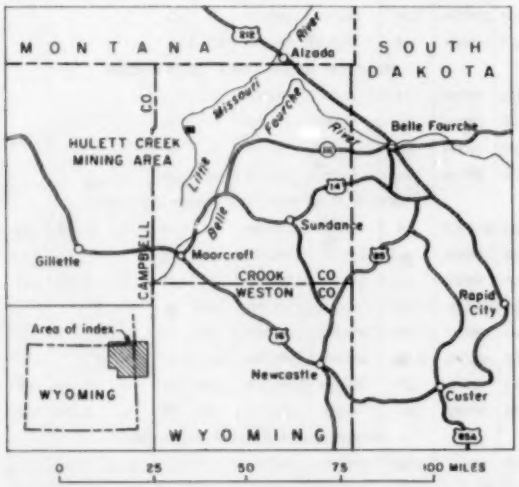



FIG. 1. Map showing location of Hulett Creek area, Wyo.

Hills, or elsewhere, although results of other radiochemical analyses indicate that many of the uranium deposits in Wyoming have a similar range in age to those of the Hulett Creek area.

## GENERAL GEOLOGY

The Hulett Creek area is on the northwestern flank of the Black Hills in Crook County, northeastern Wyoming (fig. 1). It is an area of gently rolling topography cut by comparatively steep-walled stream valleys. The maximum relief, excluding the surrounding hills is about 200 feet.

The rocks exposed in the Hulett Creek area (7) include in ascending order the Lakota formation, Fall River formation, and Skull Creek shale, all of Early Cretaceous age. The Fall River formation, which consists of fine- to medium-grained sandstone and interbedded siltstone and shale, essen-

tially forms the topographic surface of the area. The Lakota formation, which unconformably underlies the Fall River formation, consists of fine-grained to conglomerate sandstone and interbedded claystone. It is exposed only in the deeper stream valleys and in the Hauber mine. The Skull Creek shale, which gradationally overlies the Fall River formation, crops out in isolated patches on the Fall River surface and forms the hills marginal to the district. A generalized section of these formations and the underlying Morrison formation, of Late Jurassic age, is given in Figure 2.

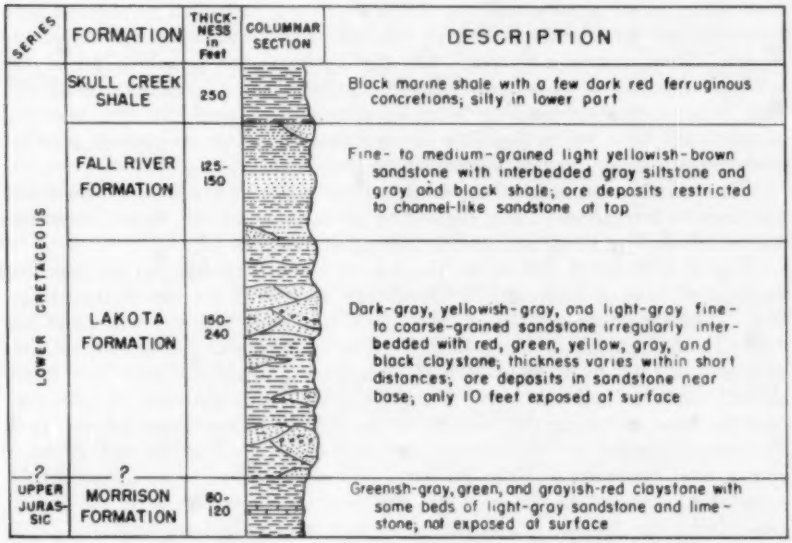

SERIES
FORMATION
THICKNESS in Feet
COLUMNAR SECTION
DESCRIPTION
LOWER CRETACEOUS
SKULL CREEK SHALE
250
Black marine shale with a few dark red ferruginous concretions; silty in lower part
FALL RIVER FORMATION
125-150
Fine- to medium-grained light yellowish-brown sandstone with interbedded gray siltstone and gray and black shale; ore deposits restricted to channel-like sandstone at top
LAKOTA FORMATION
150-240
Dark-gray, yellowish-gray, and light-gray fine- to coarse-grained sandstone irregularly interbedded with red, green, yellow, gray, and black claystone; thickness varies within short distances; ore deposits in sandstone near base; only 10 feet exposed at surface
?
?
UPPER JURASSIC
MORRISON FORMATION
80-120
Greenish-gray, green, and grayish-red claystone with some beds of light-gray sandstone and limestone; not exposed at surface

FIG. 2. Generalized section of the formations of the Hulett Creek area, Wyo.

On the divides between the stream valleys are local deposits of gravel and loess, and within the stream valleys four terrace levels of alluvium. The gravel deposits on the divides are considered to be of possibly Pleistocene age or older, and the loess deposits of late Wisconsin age or younger (H. D. Goode, written communication, 1956). The higher alluvial deposits in the valleys might be as old as late Wisconsin.

Structurally, the Hulett Creek area is a terrace on the east side of the Black Hills monocline. Dips within the area range from 2° to 5° N and NW with local reversals of 1° to 2°. Along the Black Hills monocline at the west edge of the area the dips steepen abruptly to 10° to 90° W and NW. The Black Hills monocline is defined as the boundary between the Black Hills uplift on the east and Powder River basin to the west.

A series of normal faults, ranging in length from about 10 feet to a mile, cut northeastward from the Black Hills monocline across the Hulett Creek area. The faults strike N 45° to 60° E and dip 50° to 90° NW or SE.

Along most of the faults the northwest side is downthrown, the throw ranging from less than a foot to 60 feet. Regional studies show that the folding was pre-Oligocene and that the faulting in the Hulett Creek area is probably the result of the formation of the Black Hills monocline.

Ground water in the Hulett Creek area is controlled by the lithology and structure of the Fall River and Lakota formations. In drilling, as far as the authors know, all the rocks at 150 feet or more below the surface were saturated with water. In the upper 150 feet, to within about 10 feet of the surface, many sandstone lenses are saturated with ground water. The controlling factor as to whether a sandstone lens is saturated or not has been its shape and structural position in relation to the downward cutting of the Hulett Creek drainage system. On the basis of the possible age of the surficial deposits, it is believed that the ground-water level was above the Fall River formation until at least Wisconsin time, and that the lowering of the water table below the Fall River formation, or to its present position occurred principally in late Wisconsin to Recent time.

The uranium deposits of the Hulett Creek district are in the Fall River and Lakota formations. The deposition of the uranium in these formations was controlled by structural and stratigraphic features (5, 8).

In the Fall River formation, the uranium was deposited at or near the margins or base of a channel-like sandstone deposit at the top of the formation. The sandstone deposit is about 2,000 feet long, 800 feet wide, and has a maximum thickness of about 45 feet. The main part of the deposit consists of lenses of fine- to medium-grained crossbedded sandstone with thin interbedded siltstone and claystone seams. Carbonaceous material is abundant near the base and along the margins of the individual sandstone lenses where the deposit grades to very fine grained thin-bedded sandstone and siltstone with a total thickness of 10 to 20 feet. The sandstone deposit trends northwest about parallel to the regional dip and at right angles to the faults. Underlying the deposit is a sequence of thin interbedded siltstone and shale from about 5 to 30 feet thick, the thickness varying inversely as the thickness of the sandstone deposit.

A structure-contour map on the base of the channel-like sandstone shows that the uranium deposits occur in depressions, which are the result of deposition of the sandstone and subsequent folding and faulting, and at the margins of the sandstone deposit where facies changes have been subsequently folded and faulted. The uranium was deposited within this channel-like sandstone where the movement of the ore solutions was constricted by structural and stratigraphic features.

The ore deposits of the Fall River formation may be grouped, on the basis of their relation to a local water table, into those well above any local water table and those just above or just below a local water table.

The ore deposits in the Lakota formation are in sandstone lenses, ranging in thickness from 2 feet or less to 30 feet, near the base of the formation, about 350 to 400 feet below the topographic surface and about 250 feet below the top of the zone of ground-water saturation. The mineralized sandstone lenses occur on either side of a northwest trending fault, which

has a maximum displacement of about 30 feet at the surface, with the northeast side downdropped.

## CHEMICAL ANALYSES

Table 1 lists the grade (percent uranium) and age calculations, based on radiochemical analyses, of samples from the Hulett Creek area. This table includes, in addition to the results calculated in part I of this paper, the results previously calculated (9) for deposits below the water table. A limited number of elements were determined by chemical methods. Semiquanti-

TABLE 2. ANALYSES AND ATOMIC RATIOS OF VARIOUS ELEMENTS TO URANIUM OF SAMPLES FROM THE HULETT CREEK AREA, WYOMING

| LABORATORY NUMBER | FIELD NUMBER | U % BY WT. | V % BY WT. | V / U | FE % BY WT. | FE / U | CA % BY WT. | CA / U | MG % BY WT. | MG / U | MN % BY WT. | MN / U | AS PPM | AS / U x10⁻⁴ | SE PPM | SE / U x10⁻⁴ |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | | DEPOSITS ABOVE THE WATER TABLE (CARNOTITE-TYUYAMUNITE DEPOSITS) | | | | | | | | | | |
| 252409 | 79CR56 | 2.11 | 0.084 | 0.19 | 3. * | 6.1 | 1.5 * | 4.2 | 0.3 * | 1.4 | 0.07 * | 0.14 | 120 | 1.8 | 900. | 12.8 |
| 252410 | 80CR56 | .34 | .14 | 1.9 | .3* | 3.8 | 3. * | 52.4 | .7 * | 20.2 | .3 * | 3.7 | 14 | 1.3 | 10. | .9 |
| 252411 | 81CR56 | .40 | .17 | 2.0 | .7* | 7.5 | .15* | 2.2 | .03* | .7 | .3 * | 3.1 | 32 | 2.5 | 50 | 3.8 |
| 252412 | 82CR56 | .38 | .13 | 1.6 | .3* | 3.4 | .15* | 2.3 | .03* | .8 | .15 * | 1.6 | 16 | 1.3 | 5 | 2.4 |
| 252418 | 112CR56 | 3.06 | .72 | 1.1 | 1.5* | 2.1 | 3. * | 5.8 | .15* | .5 | .07 * | .1 | 174 | 1.8 | 5 | .05 |
| 252419 | 113CR56 | .69 | .28 | 1.9 | .3* | 1.9 | .7 * | 6.0 | .15* | 2.1 | .07 * | .4 | 140 | 6.4 | 4 | .17 |
| | | | | | | DEPOSITS AT THE WATER TABLE (PYRITE OXIDIZED) | | | | | | | | | | |
| 252424 | 87CR56 | 0.13 | 0.015* | 0.54 | 2.48 | 81.5 | 2.57 | 118. | 0.03* | 2.3 | 0.164 | 5.5 | 204 | 90. | 8 | 1.9 |
| 252426 | 89CR56 | .010 | .007* | 3.3 | 1.80 | 769. | .43 | 255. | .03* | 29. | .012 | 5.2 | 58 | 185. | 200 | 604. |
| 252430 | 93CR56 | .034 | .015* | 2.06 | 2.62 | 328. | .86 | 150. | .07* | 20. | .012 | 1.5 | 196 | 183. | 10 | 8.9 |
| 252432 | 97CR56 | .016 | .007* | 2.04 | 1.15 | 306. | 2.51 | 933. | .03* | 18. | .284 | 77. | 76 | 151. | 2 | 378. |
| | | | | | | DEPOSITS AT THE WATER TABLE (PYRITE UNOXIDIZED) | | | | | | | | | | |
| 252423 | 86CR56 | 0.25 | 0.015* | 0.28 | 1.65 | 28.2 | 0.07* | 1.7 | 0.18 | 7.1 | 0.012 | 0.21 | 58 | 74. | 1 | 0.12 |
| 252425 | 88CR56 | .043 | .015* | 1.6 | 1.85 | 183. | .07* | 9.7 | .06 | 14. | .015 | 1.5 | 78 | 378. | 4 | 2.8 |
| 252427 | 90CR56 | .45 | .003* | .03 | 1.54 | 14.6 | .14 | 1.8 | .06 | 1.3 | .012 | 1.2 | 58 | 41. | 40 | 2.7 |
| 252428 | 91CR56 | .25 | .007* | .13 | 2.29 | 39. | .14 | 3.3 | .18 | 7.1 | .011 | 1.9 | 40 | 51. | 4 | .48 |
| 252429 | 92CR56 | .16 | .007* | .20 | 4.51 | 120. | 3.72 | 138. | .12 | 7.4 | .557 | 15. | 22 | 44. | 3 | .56 |
| 252431 | 94CR56 | 1.56 | .007* | .02 | 3.32 | 9.1 | .14 | .53 | .03* | .19 | .010 | .28 | 875 | 18. | 100 | 1.9 |
| 252433 | 96CR56 | .18 | .007* | .18 | .48 | 11.4 | 3.43 | 113. | .18 | 9.8 | .360 | 8.7 | 30 | 53. | 1 | 17. |
| 252434 | 99CR56 | .12 | .007* | .27 | .80 | 28.4 | .93 | 46. | .03* | 2.5 | .022 | .80 | 70 | 19. | 3 | 75. |
| | | | | | | DEPOSITS BELOW WATER TABLE (HAUBER MINE) | | | | | | | | | | |
| 269010 | H1 | 0.048 | 0.025 | 2.4 | 0.13 | 11.5 | 0.15 | 18.5 | 0.078 | 16. | $7.2\times10^{-5}$ | $6.5\times10^{-3}$ | 11 | 73. | 0.5 | 0.3 |
| 269011 | H2 | .33 | 1.01 | 14.3 | .41 | 5.3 | .56 | 10.1 | .31 | 9.2 | $14.8\times10^{-5}$ | $1.9\times10^{-3}$ | 90 | 8.7 | 90 | 4.6 |
| 269012 | H3 | .56 | .57 | 4.7 | .32 | 2.4 | 1.05 | 1.1 | .13 | 2.3 | $7.6\times10^{-5}$ | $5.9\times10^{-3}$ | 37 | 2.1 | 9 | .5 |
| 269013 | H4 | 1.30 | .11 | .40 | .12 | .39 | 1.71 | 7.8 | .14 | 1.1 | 72 $\times10^{-5}$ | 24 $\times10^{-3}$ | 16 | 3.9 | 8 | .2 |
| | MILL PULP LOT NUMBERS | | | | | | | | | | | | | | | |
| 271915 | 188 | .17 | .17 | 4.6 | 1.96 | 49. | 3.02 | 105. | .90 | 52. | .05 | 1.3 | 65 | 12. | 40 | 7.1 |
| 271916 | 189 | .19 | .17 | 4.2 | 1.40 | 31. | 2.93 | 92. | .90 | 46. | .07 | 1.6 | 65 | 11. | 35 | 5.6 |
| 271917 | 190-191 | .16 | .17 | 5.0 | 2.38 | 64. | 3.18 | 118. | .90 | 55. | .07 | 1.9 | 65 | 13. | 50 | 9.4 |
| 271918 | 192-193 | .17 | .17 | 4.7 | 1.54 | 39. | 3.08 | 108. | 1.02 | 59. | .08 | 2.0 | 65 | 12. | 50 | 8.9 |

* SEMIQUANTITATIVE SPECTROGRAPHIC ANALYSES: LIMITS OF DETECTION, IN PERCENT; V, 0.001; FE, 0.0008; CA, 0.005; MG, 0.0005; MN, 0.0002. ANALYSTS: J.C. HAMILTON AND R.G. HAVENS.

ANALYSTS: U, R.J. FENNELLY, H.H. LIPP, AND D.L. FERGUSON; V, W.D. GOSS AND J.S. WAHLBERG; FE AND MN, W. D. GOSS AND D. L. SKINNER; CA AND MG, W. MOUNTJOY AND R.J. FENNELLY; AS, J.P.SCHUCH, C. HUFFMAN, AND D.L. FERGUSON; SE, G.T. MURROW.

tative spectrographic analyses were made on all samples. The results of these analyses are given in Table 2.

The results of the radiochemical and other analyses of Tables 1 and 2 are listed in order from deposits above the water table to those below. The analyses from the deposits at the water table are further classified as to whether the sample contained pyrite, or the pyrite had been oxidized to

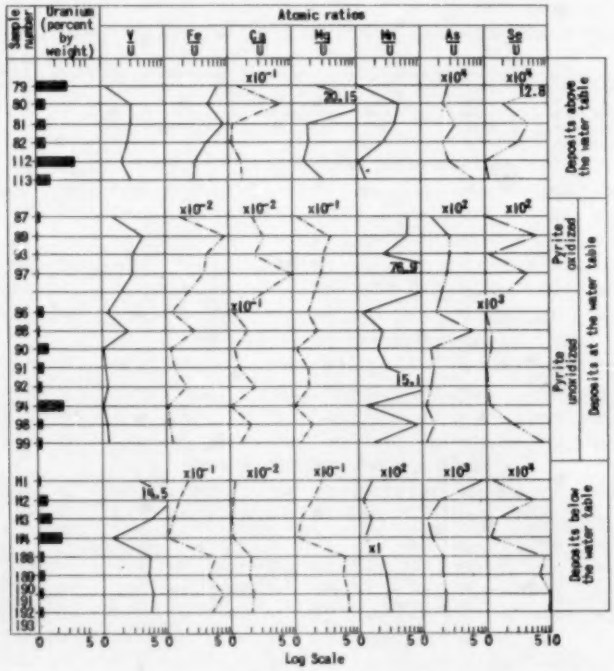



FIG. 3. Graphs of atomic ratios of various elements to uranium of samples from the Hulett Creek area, Wyo.

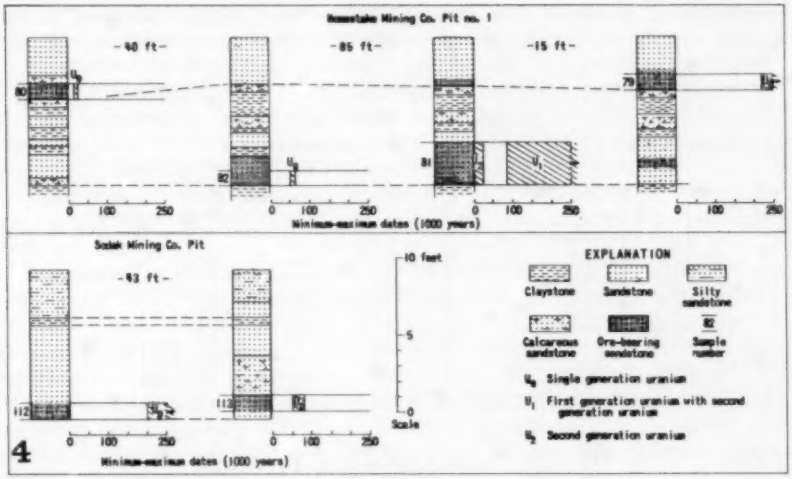
iron oxides. The criterion was useful in the detailed mapping and sampling of the deposits as they were exposed by mining, and correspond to whether the sandstones were saturated or not with water.

Table 2 also gives the atomic ratios of the various elements to uranium and the results are plotted on Figure 3. These results illustrate the differences in the atomic ratios for the different geologic conditions.

## URANIUM MIGRATION DATES RELATED TO GEOLOGY AND ANALYTICAL RESULTS

Geologic and chemical factors influence the deposition and migration of the uranium and other elements. The data are not complete enough to

establish all the controlling factors, but certain relations between the geology, the apparent age, and the chemistry can be interpreted from the results to date. Figures 4, 5, and 6 illustrate the geologic environment from which the samples were taken together with the graphs of the apparent minimum and maximum dates of uranium introduction.



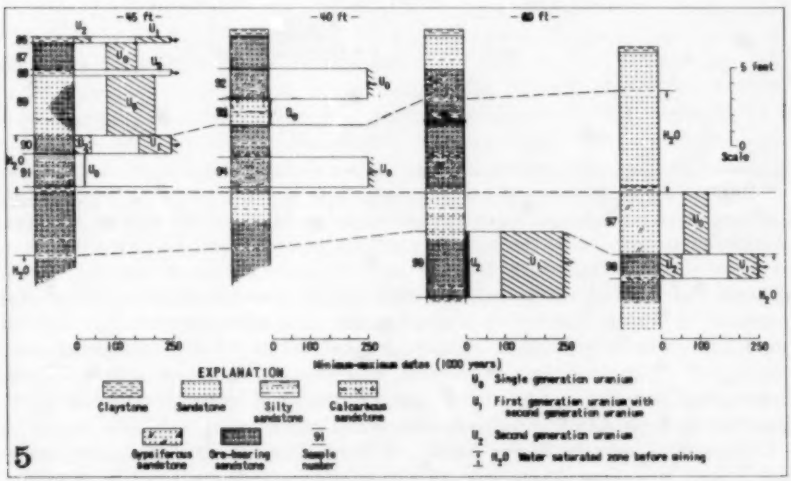




FIG. 4. Geologic environment of samples from deposits above the water table, and graphs of the apparent minimum and maximum dates of uranium introduction.

FIG. 5. Geologic environment of samples from deposits at the water table, and graphs of the apparent minimum and maximum dates of uranium introduction.

To interpret the radiochemical results, different types of uranium must be implied. For this discussion, the term "original uranium" applies to the uranium in those deposits that have only a single generation of uranium, regardless of the age of the deposits. The term "first generation uranium" applies to the older uranium and the term "second generation uranium" applies to the younger uranium in those deposits that contain more than one generation of uranium. As will be shown, the "original uranium" of a deposit may be equivalent in age to the "first generation uranium" of one

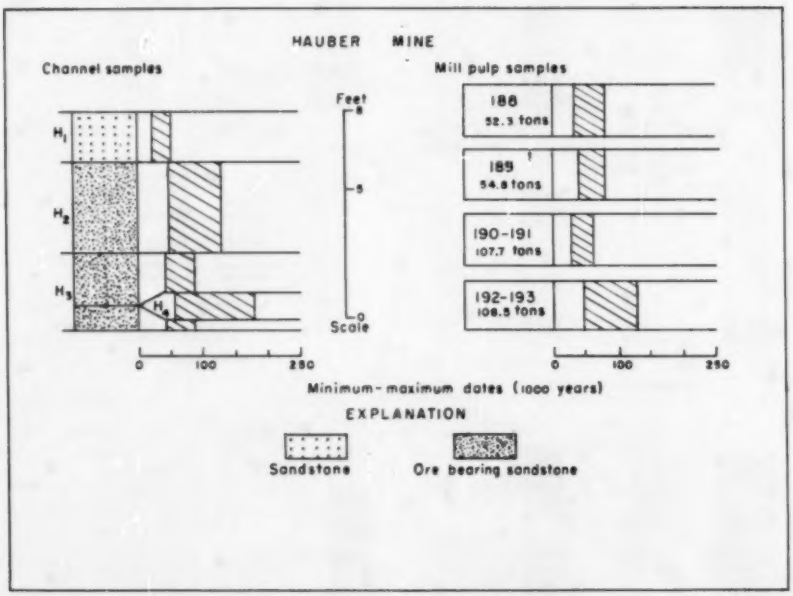



FIG. 6. Geologic environment of samples from deposits below the water table, and graphs of the apparent minimum and maximum dates of the start of uranium accumulation.

deposit and the "second generation uranium" of another deposit. "Primary uranium" refers to the initial mineralization in a particular locality, and in some cases can be equivalent to the "original uranium" or "first generation uranium." "Original, first generation, and second generation uranium" are determined from the results of the age calculations, where either one or more than one cycle of uranium introduction is indicated.

*Deposits Above the Water Table.* The ore deposits above the water table are typical carnotite-tyuyamunite sandstone uranium deposits. The uranium minerals, principally tyuyamunite, occur as disseminated grains or cement in the sandstone in irregularly shaped patches and streaks, and in and around fragments of carbonaceous material. An ore body, which consists of a

concentration of patches and streaks, is in general tabular parallel to the bedding. The contact of an ore body with the surrounding barren rock generally is gradational but may be sharp. Associated with ore deposits are concretions of sandstone cemented with calcium and magnesium carbonate—many several feet in maximum dimension—and concretions cemented with iron oxides—generally less than 1 inch in diameter. Pyrite, though rare, may be associated with these deposits as disseminated grains in the carbonate-cemented concretions or in the carbonaceous material.

Carnotite-tyuyamunite samples were taken from two deposits, the Homestake Mining Company pit 1 and the Sodak Mining Company pit. In pit 1 the ore occurred in two distinct bands separated by a ledge of carbonate-cemented sandstone and a claystone lens. The upper band of ore lay directly above the carbonate-cemented sandstone. In the lower band the uranium was concentrated principally near the base just above a claystone lens at the top of the siltstone and shale unit underlying the channel-like sandstone deposit (Fig. 4).

The uranium in samples 79 and 112 have similar ranges in apparent age from about 200,000 to greater than 250,000 years, and sample 81 indicates an older uranium component. Samples 82 and 113 have uranium of similar ranges in apparent age from about 54,000 to 84,000 years, and sample 81 indicates a component in this range. The uranium in sample 80 and the second generation of uranium in sample 81 have similar ranges in apparent age from 0 to 25,000 years when it is considered that, for the second generation enriched samples based on the assumptions used (page 1399), the minimum age is zero.

The older deposits, as represented by samples 79 and 112, contain, for those deposits above the water table, anomalously high values of iron (Table 1). Sample 81, which contains some first generation uranium, has an intermediate percentage of iron. Samples 80, 82, and 113, which contain only younger uranium, have a low percentage of iron. This might be explained by the fact that in the deeper deposits it has been noted that the uranium minerals are precipitated as a very fine grained mixtude with pyrite, and that the primary uranium is deposited in sandstone high in carbonaceous material, which contains abundant syngenetic pyrite. On oxidation, the uranium is soluble, but according to the work of Hem and Cropper (6) ferric iron in aerated natural ground waters of a pH greater than five can only be in solution in amounts less than 0.01 ppm, and the oxidation of ferrous to ferric iron is rapid in the normal range of pH for natural waters. Uranium on oxidation could migrate in the ground water away from the deposits, but the iron would precipitate in situ.

Garrels and Christ (3) point out that the primary uranium minerals are relatively insoluble under reducing conditions and that the vanadates, carnotite and tyuyamunite, are relatively insoluble in an oxidized environment. Much of the uranium, which oxidizes before the vanadium, could have been removed in solution before it was fixed as vanadates. With the lowering of the general water level below the the deposits, and the draining of any perched water tables, the uranium deposits would be oxidized to carnotite

and tyuyamunite, and the deposits would be relatively stable, if enough vanadium was present to form vanadates. In these deposits there is an excess of vanadium for the formation of stable vanadates except in the deposit represented by sample 79 (Table 2). Sample 79 contained abundant carbonaceous material, and according to studies of Breger and Deul (1), mechanisms are known that would permit the retention of uranium by coal, coalified wood, or related substances regardless of the state of oxidation of the uranium in solution.

From these results, and the relative positions of the deposits, it could be assumed that the primary mineralization occurred more than 250,000 years ago and that the younger uranium represents migration and repreciptation of the primary uranium. The oxidation of the primary uranium would depend on the lowering of the ground-water table to or below the deposits. Samples 79 to 112 indicate that the primary ore was oxidized in place; however, there may have been a small amount of migration of the primary uranium present. Sample 81 would indicate that some uranium, probably from a primary deposit stratigraphically and structurally higher, was added to the first generation uranium. Samples 80, 82, and 113 probably represent uranium deposits displaced by migration a significant distance from the primary deposits before being fixed by vanadates. The migrated deposits, which provide the closest indication of the date of oxidation, are not all of the same age, but all are less than about 80,000 years old. The time and place of precipitation of the migrated uranium would depend on the direction of flow of the uranium bearing ground water, which would change from time to time in such lenticular stratigraphic units, as the lowering of the general ground-water level progressed.

The minimum apparent age for a deposit above the water table, excluding sample 81, for which the minimum age of the second generation uranium can not be determined, is 15,000 years. From this it could be assumed that most of the uranium deposits were oxidized and stabilized by the formation of tyuyamunite and carnotite by this time. This date agrees with the geologic evidence that indicates that the water level was lowered below pit 1 and the Sodak pit by late Wisconsin or early Recent time, which would be, interpreting the C-14 dates as reported by Flint (2, p. 326), between 15,000 and 8,000 years before present. It can be assumed then, that oxidation and second generation mineral deposition in these deposits occurred less than about 80,000 years ago and may have extended over most of the period between then and about 10,000 years ago.

*Deposits at the Water Table.* The samples for deposit at the water table were taken from the Homestake Mining ompany pit 5. In this deposit the uranium occurred in a series of sandstone lenses at the base of the upper sandstone unit of the Fall River formation, which has been faulted and folded to form a structural basin. In the structural basin were two superimposed perched water tables. The samples were taken from both above and below the water levels (Fig. 5).

The sandstone in that part of the deposits above the water tables is commonly light yellowish brown to brown, the color resulting from iron-oxide

stain on the sand grains. Concretions cemented with gypsum, and less commonly carbonate, ranging from less than an inch to a few feet in diameter, are abundant. A few pyrite grains may be present in the concretions and the carbonaceous material, but most of the pyrite has apparently been oxidized to form the iron-oxide stain that colors the sandstone. The sandstone of that part of the deposit below the water table is light to dark gray. The coloration results from a very fine grained mixture of pyrite and opaque black uranium minerals that locally coat and embay the sand grains and cements the sand grains, and from disseminated fine-grained specks of carbonaceous material. The carbonaceous material also occurs as fragments up to an inch long concentrated in streaks along bedding planes and as seams of coal-like material up to 2 inches thick. Carbonate cemented concretions and beds are common locally, and some carbonate cement is usually associated with the ore minerals.

Efforts to identify the ore minerals in samples from the deposits at the water table in the laboratories of the U. S. Geological Survey were unsuccessful (L. B. Riley, written communication, 1958). Examination of thin and polished sections, however, and a comparison of those with others from similar uranium deposits in the northern Black Hills where the minerals have been identified by X-ray analyses, indicates that the minerals from the deposits below the water levels are probably uraninite and coffinite (R. E. Davis and G. A. Izett, written communication).

The samples for deposits at the water table were taken from the Homestake Mining Company pit 5. In comparing the chemical analyses of the samples above the water levels in pit 5 and the samples of pit 1 and the Sodak pit, the chemical composition is similar except for the percentages of uranium and vanadium, which in the samples of pit 5 are 10 to 100 times lower. The ratios of vanadium to uranium, however, are about the same. A comparison of the analytical results between the two different types of deposits in pit 5 shows that the principal difference is in the percentage of uranium, which, in general, is more than 10 times greater in the samples taken from the water saturated zones.

From the calculations of apparent ages based on the radiochemical analyses (Table 1), the minimum and maximum ages indicate the presence of a young age uranium component in all the samples from the water saturated zones with the exception of sample 94, which shows evidence of undergoing considerable uranium leaching and thus is not a favorable site for uranium enrichment. The young component appears as second generation uranium in all of these samples except for sample 91, which is considered as a special case where uranium derived from above was deposited in essentially barren sandstone. Thus the deposits below the water levels have been affected, predominantly, by processes of secondary enrichment. From the relative positions of the secondarily enriched samples, it is logical to assume that the secondary uranium was derived from oxidized deposits above the water levels and reprecipitated under reducing conditions below the water levels (Fig. 5).

From the deposits above the water levels, only sample 86 indicates second-

ary enrichment of a magnitude comparable to those below the water levels. This sample is from a relatively impervious clay seam that underlay a weakly mineralized sandstone that was stripped prior to mining. Sample 88 is also a relatively impervious claystone and sample 92 is a relatively impervious gypsum cemented concretion. Of the samples above the water levels, only these three show the existence of visible pyrite and the presence of uranium which could be much greater than 250,000 years old. All of the other samples taken from above the water levels (87, 89, 93 and 97) occur in permeable sandstone, contain, generally, much less uranium, no visible pyrite, and indicate no significant uranium component greater than 250,000 years old. The permeable sandstone deposits analyzed all show a trail of radioactive daughter products considerably in excess of that which could be supported by the uranium content. This suggests that these sandstones were simply pathways for the migration of relatively large amounts of uranium moving downward from leached deposits above. Only a small amount of uranium remains fixed in the sandstone, while the bulk of uranium that would be required to support the daughter products, represents uranium in transit. If uranium was in transit in these samples, it probably had not remained in the host rock for a time that exceeds one-half of the total time that uranium has been migrating through this zone, as required by the second assumption for calculating the apparent age (page 1395). In this case, the calculated apparent minimum age would be too old, and the actual date of the beginning of significant uranium migration through the zone would be less than the apparent minimum age.

A deviation from the first assumption for calculating the apparent age (page 1395), that the daughter products between uranium and radium do not migrate in measurable amount, is indicated for sample 93. An excess of $Pa^{231}$ compared to $Th^{230}$, above the theoretically possible excess for the case of no daughter product migration, is evident for this sample. The fact that such a deviation does occur indicates the presence of significant volumes of ground water flowing through this zone in recent times.

The calculations of both the minimum and maximum dates for samples 88, 92, and 94 indicate that the uranium in these three locations is much greater than 250,000 years old. From this it is concluded that the age of the first generation uranium in the water saturated zone is also much greater than 250,000 years. The older uranium in the samples from the water saturated zones then represents primary deposits to which secondary uranium has been added. Assuming that the second generation uranium represents secondary enrichment derived from a source of oxidizing uranium above the water level, the age relationships indicate that the enrichment had reached the level sampled during the last 60,000 years. Also the enrichment process has operated at several intervals, or continuously, during the past 60,000 years. This would confirm the observations made for the deposits above the water table, where it was concluded that the deposits less than 80,000 years old were derived from older deposits, and the oxidation and migration of the uranium started some 80,000 years ago.

*Deposits Below the Water Table.* All the samples, as referred to

below the water table, are from the Hauber mine and from at least 250 feet below the top of the zone of ground water saturation. Three of the samples were taken across an 8-foot ore lens, the fourth is a high-grade sample from the same lens. The four additional samples are from mill pulps and represent about 300 tons of ore (Fig. 6).

The uranium mineral, or minerals, based on polished- and thin-section examinations, are presumed to be uranite and coffinite, and occur as very fine grained opaque mixtures with pyrite interstitial to and as coatings on the sand grains, and as filling small veinlets that cut the gangue minerals and sand grains. Interstitial to the sand grains, where not replaced by ore minerals, are clay, carbonate and silica cement, and pyrite. The ore minerals replace, in order of preference, the clay, pyrite, and sand grains. Veinlets of calcite locally cut the ore and the gangue minerals. An ore body consists of a concentration of small spots and streaks of sandstone and carbonaceous material that have been impregnated with the mixture of ore minerals and pyrite.

The samples from the Hauber mine should have had a much simpler geochemical history than those deposits above or at the water table, which have been subject to oxidation, leaching, and secondary enrichment. A different concept for the apparent age results in samples from far below the water table was proposed by Rosholt (8). This concept defines the uranium accumulation as a possible continuous process, beginning somewhere between the maximum and minimum date, and proceeding to, or nearly to, the present time. The range in apparent dates for the start of the uranium accumulation in the Hauber mine is 23,000 to 180,000 years ago. The highest grade sample (H4) indicates that this uranium accumulation began somewhere between 60,000 and 180,000 years ago. The results for the four mill pulp samples, representing about 300 tons of 0.17 percent uranium ore, indicate that the earliest significant uranium accumulation in this deposit did not start before about 130,000 years ago.

It should be noted that sample H4 is a selected high grade sample taken from 1 foot above the base of channel sample H3. The difference in the apparent age results for samples H3 and H4, where sample H3 actually contains the interval of sample H4, illustrates the effect of sampling on the results. For this reason, samples for this study were selected that had been taken completely across a stratigraphic unit and therefore should be representative of the whole unit.

In comparing the analyses of the samples from the Hauber mine, with the other deposits, the most striking difference is in the vanadium. The carnotite-tyuyamunite deposits have about the same percentage of vanadium, but the ratio of vanadium to uranium is only about 2:1 whereas in the Hauber mine the ratio is about 4:1. As compared to the deposits at the water table, the percentage of vanadium in the Hauber mine is about 10 times as great. The abundance of the other elements are essentially the same between the types of deposits, except for manganese from the channel samples from the Hauber mine, which is very low. The percentages of arsenic and selenium in the Hauber mine samples are within the range of the percentages of these

elements in the other deposits, and the values are more consistent from sample to sample.

From Figure 3 it may be seen that in the Hauber mine the distribution of the atomic ratios of iron, calcium, magnesium, manganese, and arsenic to the uranium are similar and that of the vanadium and selenium ratios with uranium are similar. Of interest, is the offset of the graphs between samples H4 and 188 (Fig. 3). This offset represents the effects of mining and the mixing of ore to maintain a constant grade. The extreme cases are the ratio of manganese to uranium, which decreased almost 100 times, and the ratio of arsenic to uranium, which is not appreciably effected.

## SUMMARY

The results presented represent samples from three distinct geologic and chemical environments through a stratigraphic range of 400 feet within a horizontal distance of 4,000 feet.

The geology of the area indicates that the deposits were formed by ground water migrating down dip in channel-like sandstone deposits. The deposits were formed in chemically favorable environments where the movement of the ground water was constricted by structural and stratigraphic traps.

The chemical analyses show that the average values of iron, calcium, magnesium, and manganese are similar, or within the probable sampling error, regardless of the environment from which the samples were taken. The values of arsenic and selenium are less consistent in the samples from the deposits at and above the water table. Vanadium values for the samples from the deposits above the water table and from below the water table are similar, and more than 10 times the values in samples from the deposits at the water table. The values for uranium reflect the type of deposit and the geologic position of the sample within the deposit, as illustrated best by the channel samples from the deposits at the water level.

Figure 3 illustrates the variation in the relative amounts of various elements to uranium, and because the uranium is used as a common denominator, to each other. All the elements do not form distinct mineral species with the uranium, but all are effected to varying degrees by the chemical conditions that influence the precipitation and migration of uranium. This illustration shows that there is no apparent correlation between the distribution of the ratios for the deposits above the water table, that the distribution of the ratios for vanadium, iron, and arsenic are similar for the deposits at the water table, and that the distribution of the ratios of iron, calcium, magnesium, manganese, and arsenic are similar and the ratios of vanadium and selenium are similar in the deposits well below the water table. From this it can be concluded that the chemical history of the carnotite-tyuyamunite deposits is the most complex, for the deposits at the water table somewhat less complex, and for the deposits well below the water table, relatively simple.

Uranium deposits were formed in the higher stratigraphic units more than 250,000 years ago. Oxidation, and the alteration of the primary uranium

minerals to carnotite and tyuyamunite, and the migration of uranium and the secondary enrichment of the deposits has taken place in the last 80,000 years. For those deposits above the water table, the migration of the uranium largely stopped about 10,000 to 15,000 years ago. For those deposits at the water table, the oxidation, migration, and secondary enrichment of the older deposits has continued to the present time. In the lower stratigraphic units uranium deposition did not start earlier than about 180,000 B.P. and has continued probably to about the present, in part nearly contemporaneously with much of the oxidation and secondary enrichment of the older deposits nearer the present surface.

The conclusions reached by the authors, based on this preliminary study, is that if the geology and chemistry of a uranium deposit in sandstone are known, whether the deposit is above, at, or below the water table, it is possible with radiochemical analyses to determine the time of introduction or migration of the uranium, if the time is less than 250,000 years B.P.

## APPLICATION OF RADIOCHEMICAL ANALYSES AND URANIUM MIGRATION DATES IN PROSPECTING

From the radiochemical analyses, and by the same methods used to calculate the apparent age of uranium introduction, it is possible to calculate the probable grade of a leached deposit (Table 1). Additional studies are needed to evaluate the accuracy of these apparent ages and grades, or the chemical factors that might influence the results, but it seems reasonable to assume that with additional studies a relatively accurate range in age and original grade for a leached deposit could be determined.

In the northern Black Hills are several areas of anomalous radioactivity that would indicate commercial deposits of ore-grade material at shallow depths. Exploration of these areas has shown, however, no commercial ore deposits. It should be possible for such areas to determine with radiochemical analysis the original grade—and by mapping—the original size of such deposits. This information, plus a knowledge of the stratigraphy, structure, and geomorphic history of the area, should allow the reasonable interpretation of the course of the migrating uranium and the possible sites of redeposition, which might be at such a depth, as is the Hauber mine, that the deposit could not be detected by anomalous radioactivity from the surface. It should also allow a reasonable guess to be made as to whether such a deposit would be of commercial size and grade.

## ACKNOWLEDGMENTS

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# CROCIDOLITE FOM THE KOEGAS-WESTERBERG AREA, SOUTH AFRICA

J. J. LE R. CILLIERS, A. G. FREEMAN, A. HODGSON, AND H. F. W. TAYLOR

ABSTRACT

The geology of the Koegas area is described, with particular reference to the occurrence of crocidolite (fibrous riebeckite). The various crocidolite-bearing reefs are listed, and the weathering of the fiber and its associated rocks is discussed. Twenty-two samples of crocidolite from precisely defined locations in both Fresh and Weathered Zones have been examined, mainly by chemical analysis, X-ray diffraction, and differential thermal analysis; these samples are collectively representative of a large amount of workable ore. The chemical composition of the fiber from the Fresh Zones is almost constant, and approximates to $Na_2Fe'''_2Fe''_{2.5}Mg_{0.5}Si_8O_{22}(OH)_2$, i.e. riebeckite with one-sixth replacement of ferrous iron by magnesium. The unit-cell parameters, referred to body-centered axes, are $a$ 9.97, $b$ 17.85, $c$ 5.30 Å, $\beta$ 107°30′ ($a \sin \beta$ 9.52 Å). The observed density is 3.34 to 3.37 g/cc for most of the samples. The DTA curve, determined in air, shows exothermic peaks at 410 to 420° and 920 to 960° C and an endothermic peak at 895 to 910° C. Fibers from the Weathered Zone differ only slightly in chemical composition from the fresh material, chiefly in showing a higher ratio of ferric to ferrous iron. It does not differ appreciably in unit cell size, density, thermal behavior, or optical properties.

## PART 1. OCCURRENCE. BY J. J. LE R. CILLIERS

### INTRODUCTION

In the Cape Province deposits of crocidolite are sporadically developed in a succession of Precambrian banded ironstone beds. These rocks are exposed in a nearly continuous belt, some 320 miles long and up to 45 miles wide, which extends from 25 miles south of Prieska to the border of the Bechuanaland Protectorate in the North (Fig. 1, inset). The banded ironstone is more resistant to weathering from the underlying and overlying rocks, and generally forms hills rising to over 1,000 feet above the surrounding country.

The Koegas Asbestos Mine is situated in the Cape asbestos belt, 60 miles from its southernmost tip. The mine workings lie on either side of the Orange River, which here flows about 3,000 feet above mean sea-level. The annual rainfall is of the order of nine inches. The summer months are hot, and temperatures in excess of 100° F (38° C) are common. The winter months are cool with some frost.

### GENERAL GEOLOGY

In the Cape Province, the asbestos-bearing banded ironstone beds occupy the lower half of the Lower Griquatown Stage of the Pretoria Series, which

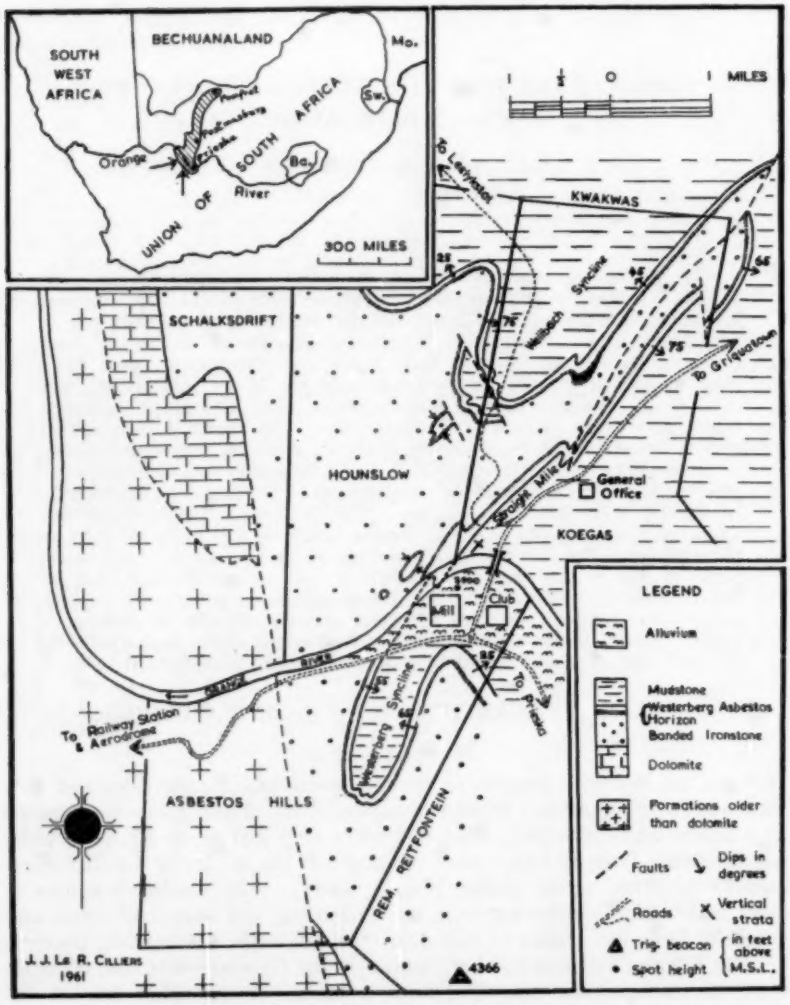


FIG. 1. Generalized geological map of the Koegas mining area. Inset (top, left): map of South Africa, showing location of the Cape Asbestos Fields (shaded). Arrows point to the area covered by the main map.

is the uppermost division of the Transvaal System. The thickness of these beds varies from 2,500 feet in the south to less than 1,000 feet in the north. Along the eastern side of the Cape asbestos belt the rocks dip regularly westwards at angles seldom exceeding 10 degrees. However, along the western side of the belt the rocks have been folded at least twice. The first

period of folding was relatively gentle and took place before the deposition of the rocks of the overlying Loskop and Waterberg Systems. After their deposition, these rocks, as well as those of the underlying Transvaal System, were subjected to more intense pressure directed from the west and they were intensely folded along the western side of the asbestos belt. In the area of maximum pressure, near Postmasburg, dragfolds were developed near the base of the banded ironstone succession where these rocks were thrust over the underlying rocks along a series of low-angle thrust faults. In the Koegas-Prieska area a series of north-plunging synclines and anticlines was formed and thrust faulting took place along the Doornberg Fault, as a result of which the banded ironstone had been eliminated west of this fault (Fig. 1). Near-vertical normal and reverse faults cut through the intensely folded banded ironstone in the Koegas mining area. Dikes have since intruded along many of these faults.

*Weathering.* The asbestos-bearing banded ironstone is extremely susceptible to alteration under the normal agencies of weathering and has undergone extensive leaching and oxidation down to depths of 250 feet below surface. Three zones of weathering are recognized. The Fresh Zone lies below the present water-table, which is generally between 50 and 200 feet below surface. In this zone the rock is finely laminated, hard, compact and dark blue to black with some pale gray bands. Above the water-table the rock is generally soft and yellowish brown due to oxidation and leaching of the constituent minerals: this represents the Oxidized or Leached Zone. Near surface these rocks have become hardened owing to secondary silification and carbonation. This zone is known as the Silicified Zone and generally forms only a thin crust near surface, although it may be completely absent in areas of high rainfall. Fortunately, the asbestos is more resistant to weathering than the enclosing rock and commonly persists right to surface with little obvious deterioration in quality. In fact, until about 1950 all the amphibole asbestos was obtained from the Leached and Silicified Zones, collectively known as the Weathered Zone. Under extreme weathering, crocidolite is oxidized to its soft, yellow pseudomorph, griqualandite. Silicification of griqualandite results in the formation of tiger's eye, which is highly prized by lapidaries.

## PETROGRAPHY OF COUNTRY ROCK

Characteristic features of the banded ironstone are its high content of magnetite and its finely laminated nature. The thickness of individual layers varies between a few centimeters and a fraction of a millimeter, the average being about five mm. The banding is generally regular, except near the base of the succession where the bands are commonly highly contorted and wrinkled. The contacts between the individual layers are sharp, and the mineralogical composition of each band remains constant, but changes completely from one band to the next.

Unweathered banded ironstone has only been exposed to any extent during the past ten years. The rock consists of alternating, thin layers of pure chert, magnetite, stilpnomelane, minnesotaite, riebeckite, and carbonates,

and layers consisting of mixtures of these minerals in which any one mineral may predominate. Pyrite and carbonaceous material are present in some layers near the base of the succession. Seams of crocidolite are interbedded in the banded ironstone along certain horizons, as described in more detail below.

The layers of pure chert consist of well-interlocking, clear grains of quartz with no preferred orientation. In the upper portion of the succession the average diameter of the grains is of the order of 0.04 mm and there is a progressive increase of grain-size with depth, so that low down in the succession the quartz grains have an average diameter of 0.08 mm. Many of the chert layers contain variable amounts of minute, slender, randomly-orientated, greenish blue needles up to 0.2 mm long, but seldom more than 0.005 mm wide. Some of the larger needles show clear pleochroism identical with that of riebeckite, but for the greater part, the needles are too fine for pleochroism to be noticeable. Some layers of chert contain disseminated euhedral grains of magnetite with an average grain size of 0.07 mm. Where amphibole needles are present in addition to the magnetite, the longer needles invariably radiate from magnetite crystals. Where a band of magnetite crystals is adjacent to a chert band, or between two chert bands that contain needles of amphibole, many of the needles have grown from the magnetite band more or less normal to the bedding-plane. The crystals of magnetite, both disseminated and in layers, must therefore have crystallized earlier than the amphibole needles, which grew preferentially from an initiating surface of magnetite where this was available.

Some of the chert bands have a speckled appearance in hand-specimen as a result of the presence of relatively large rhombs of ferruginous carbonate with major diagonals of up to 0.6 mm; these show no preferred orientation. A feature of these rhombs is that they commonly contain, generally in the centers of the crystals, patches of quartz crystals identical with those in the chert bands, together with all the accessory minerals that are present in the chert band in which they lie.

Chert bands are present that do not appear speckled in hand-specimen, but are distinctly speckled under the microscope. Under high magnification it is seen that this appearance is caused by the presence of minute laths of minnesotaite, and in places stilpnomelane, which surround the individual grains of quartz.

Bands consisting entirely of magnetite crystals are rare. The magnetite bands generally contain a small amount of quartz. Associated with this quartz are all the minerals present in the chert bands described above.

Stilpnomelane bands are abundant in the lower portion of the banded ironstone succession. The mineral grains are extremely small and are generally orientated at random. Near the base of succession the mineral is mainly of the green ferrostilpnomelane variety, whereas higher in the succession brown ferristilpnomelane predominates. The larger grains exhibit typical strong pleochroism, but as a rule the stilpnomelane bands consist of a densely packed, microcrystalline, semi-opaque mass with little visible pleochroism. Magnetite, quartz, carbonate, minnesotaite and riebeckite may

be present as accessory minerals in layers of stilpnomelane. Near the base of the succession pyrite and material of organic origin are also present in some of the layers of stilpnomelane.

Layers of minnesotaite are present throughout the beds of banded ironstone. The mineral occurs typically in microscopic sheaf-like aggregates of needles and plates that form a felt-like yellowish brown mass in which individual crystals are but rarely seen. Quartz and carbonate are common accessory minerals in these layers, and small laths of riebeckite are commonly present in greater or lesser amounts.

Some of the rock contained up to 35 percent of carbonates, largely segregated into thinly bedded layers. The carbonates approach siderite in average composition, the ratio $FeCO_3$:($CaCO_3$ + $MgCO_3$) being generally around 7.

Layers consisting entirely of riebeckite are not uncommon. Very thin layers are present throughout the banded ironstone succession, and in certain zones they become numerous and up to three feet thick. The riebeckite occurs in the form of densely matted, small laths, which are generally orientated at random or form small, fibrous spherulites. Under the microscope the effect is that of a dense, non-pleochroic mass. Where layers of massive riebeckite have been subjected to shearing, the riebeckite laths in the shear-zones have been reorientated parallel to the direction of least stress. In hand-specimen these orientated riebeckite laths show some semblance to seams of crocidolite asbestos lying at an acute angle to the bedding-plane, and are often referred to as "slip-fiber." Upon close examination it is seen that the material is brittle and completely unlike the soft, silky fibers of crocidolite.

Upon weathering the banded ironstone undergoes extensive oxidation, hydration, and silicification, so that the rock exposed on surface consists of a finely banded ferruginous and siliceous shale with intercalated thin yellow, brown and red jasper bands and a few bands of massive blue riebeckite. The rocks vary in color from pale ochre-yellow through red, purple and brown, to extremely dark brown and black, depending on the degree of oxidation and hydration of the ferruginous constituents, and the ratio of ferruginous to siliceous components. Thin, black bands of nearly pure magnetite are abundant. Color changes are sharp and clearly defined rather than gradational.

## REGIONAL METAMORPHISM AND IGNEOUS INTRUSIONS

No effects of regional metamorphism can be seen in the rocks of the Lower Griquatown Stage. Although the banded ironstone had been subjected to considerable horizontal pressure in some areas, relief of pressure took place mainly by folding. The stress set up in the rock was insufficient to generate new minerals. During the early period of folding layers of incompletely consolidated material became thicker in the crests and troughs of the folds.

A number of basic igneous sills, varying in thickness from a few feet to over 300 feet, are present in the banded ironstone. No new minerals were formed in the banded ironstone adjacent to the sills, but the grain-size of

some of the minerals, e.g. minnesotaite, is appreciably larger in the rocks adjacent to the sills. At Koegas a sill cuts through an asbestos horizon at a low angle. It was found that the amount of fiber in each reef decreased steadily as it approached the sill.

Wherever dikes cut through seams of asbestos, the crocidolite immediately adjacent to the dikes has been recrystallized to interlocking, unorientated, tabular crystals of riebeckite. This material has no commercial value and is referred to as "burnt fiber." The distance to which the crocidolite has been recrystallized on the sides of the dikes is variable, but it is generally of the order of a quarter of the width of the dike.

## CROCIDOLITE

Crocidolite asbestos is developed in the form of strictly conformable, interbedded layers along certain horizons in the banded ironstone. On the asbestos-fields a single layer of crocidolite is called a *seam* or *band* of fiber. A group of fiber bands that can be mined together as a unit is called a *reef*. A group of reefs lying near to one another, but separated by layers of barren rock of varying thickness, is called a *horizon*. An asbestos horizon occupies a specific stratigraphical position, generally has a normal thickness of between 100 and 200 feet, and consists of up to ten separate reefs each one of which contains numerous single bands of fiber.

In the Koegas-Prieska area three horizons of crocidolite are developed in the banded ironstone beds. The Lower Asbestos Horizon lies between 50 and 250 feet above the base of the succession, and all the mines in the area, with the exception of the Geduld and Koegas Asbestos Mines, extract crocidolite from this horizon. The Geduld Asbestos Mine extracts fiber from the Intermediate Asbestos Horizon, which lies approximately 700 feet above the Lower Asbestos Horizon. The Westerberg Asbestos Horizon lies about 900 feet above the Intermediate Asbestos Horizon, and about 150 feet below the top of the banded ironstone succession. It is about 150 feet thick, and it is from this horizon that crocidolite is extracted at the Koegas Asbestos Mine. The bulk of the production is obtained from the Westerberg Syncline, where fiber has now been extracted uninterruptedly for nearly seventy years, and where sufficient reserves have been established for the mine to maintain its present production for many years to come.

The distribution of the crocidolite bands in the Westerberg Asbestos Horizon is shown in a typical section in Table 1. In addition to the fiber bands longer than one eighth of an inch indicated in this table, some reefs contain up to one hundred bands of fiber narrower than one eighth of an inch, and further bands of crocidolite are developed sporadically in the waste parting between the principal reefs and above the Inner Reef.

Because longer fiber commands a higher price than short fiber, not only the total quantity of fiber in a reef, but also the length of the fiber in the individual seams is important. The length of the fibers, especially in the Lower Asbestos Horizon, commonly varies rapidly over very short distances. It is not unusual to find that the fibers in a seam of crocidolite vary in length

Table 1. Distribution of crocidolite in the Westerberg asbestos horizon

| WASTE PARTING (FEET) | REEF | REEF CHANNEL-WIDTH (inches) | AVERAGE NUMBER OF FIBRE BANDS OVER 1/8" | AVERAGE FIBRE LENGTH (inches) | MAXIMUM FIBRE LENGTH (inches) |
|---|---|---|---|---|---|
| 6 | INNER | 10 - 40 | 5 - 15 | 1/8 - 3/4 | 2 |
| 4 | MAIN | 6 - 48 | 5 - 15 | 1/8 - 2 1/4 | 4 |
| 0½ | BOTTOM MAIN | 24 - 40 | 5 - 15 | 1/8 - 1 | 1 |
| 4½ | VISSER | 48 - 82 | 5 - 18 | 1/8 - 1/2 | 1 1/4 |
| 35 | BOTTOM VISSER | 6 - 12 | 3 - 8 | 1/8 - 1/4 | 3/4 |
| 12 | INTERMEDIATE | 5 - 32 | 2 - 10 | 1/8 - 1/4 | 1/2 |
| 7 | OUTER | 9 - 48 | 5 - 18 | 1/8 - 3/4 | 1 |
| 14 | SECOND OUTER | 3 - 27 | 2 - 6 | 1/8 - 1/2 | 2 |
| | THIRD OUTER | 3 - 24 | 2 - 6 | 1/8 - 1/2 | 3/4 |

from less than an eighth of an inch to one inch over a distance of an inch or even less. This variation is typically reflected on only one surface of the seam, giving it a corrugated or cone-like aspect. Compound seams of crocidolite are common. The upper and lower surfaces of these seams are generally planar, but the seams contain one or more contorted, thin parting surfaces of non-asbestiform material, generally magnetite.

A characteristic feature of all seams of crocidolite in the Cape Province is that they are invariably capped, at least on one side, by a thin layer of

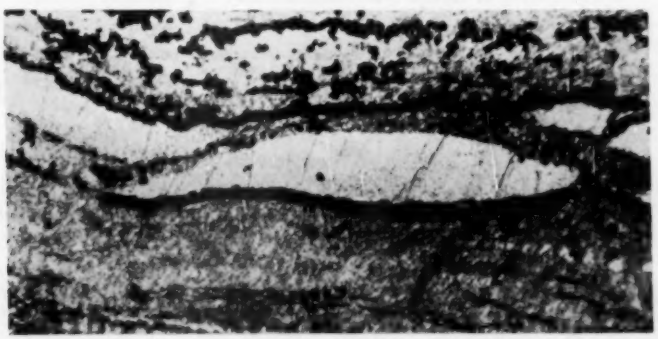

FIG. 2. Seams of crocidolite (pale gray) and magnetite (black) in massive riebeckite (medium gray, speckled). Ordinary Light, × 25. Westerberg Asbestos Horizon, Westerberg.

magnetite. The thickness of the magnetite layers is not related to the thickness of the seams of asbestos, and the layers of magnetite generally persist after the crocidolite has petered out (Fig. 2).

The latest published analyses of South African crocidolite are those included in Tables 2 and 3 of Hall's 1930 Memoir (1). These analyses show large variations in the values of $SiO_2$, $Fe_2O_3$, FeO, MgO, $Na_2O$ and $H_2O$. Specimens of unweathered crocidolite have only become available during the past ten years, and even today the crocidolite bagged at most

Table 2. Locations of Samples.

| Sample | Series, Reef, and Area | Zone | Depth below surface (feet). |
|---|---|---|---|
| RS 1 | Series 1. Main Reef, Westerberg. | Oxidised | 100 |
| 2 | | Ditto, near fresh | 167 |
| 4 | | Ditto, very near fresh | 163 |
| 5 | Series 2. Main Reef, Westerberg. | Fresh | 190 |
| 3 | | | 203 |
| 6 | | | 256 |
| 7 | | | 293 |
| 23 | Deep Borehole. Outer Reef, Westerberg. | Fresh | 1330 |
| 12 | Series 3. Outer Reef, Koegas. | Fresh | 142 |
| 14,15 | | | 166 |
| 13 | | | 186 |
| 8,9,10 | | | 236 |
| 11,16 | | | 313 |
| 18 | Series 4. Bottom Main Reef, Houmslow. | Silicified | 1 |
| 19 | | Oxidised | 71 |
| 20 | | Fresh | 148 |
| 22 | Series 5. Bottom Visser Reef, Houmslow. | Fresh | 147 |
| 21 | Inner Reef, Houmslow. | Fresh | 132 |

mines consists of a mixture of fiber obtained from the Weathered and Fresh Zones. As a result, all information quoted in the literature before 1950 refers to specimens that have been subjected to a certain degree of weathering owing to which the true character of the crocidolite has been masked to a greater or lesser extent. In view of the marked differences in the composition of the banded ironstone in the Fresh Zone as compared with that of the rock in the Weathered Zone, it was suspected that at least some of the disagreements in the published analyses may have been due to the weathering of the fiber.

In an attempt to clarify the position, five series of samples were selected from the workings of the Koegas Asbestos Mine (Table 2). Series 1

consists of samples taken from the oxidized zone of the Main Reef in the Westerberg Syncline. Series 2 consists of samples from the same reef, but only from the Fresh Zone, taken at regular intervals along the strike on two levels. Series 3 was taken from the Outer Reef in the Fresh Zone along the Koegas Straight Mile. Series 4 was taken from the Bottom Main Reef from surface down to well into the Fresh Zone on Hounslow. Series 5 was taken across the succession from 3 different reefs in the Fresh Zone on Hounslow. In addition, one specimen was taken from a deep borehole into the Outer Reef in the Westerberg Valley. It was expected that detailed examination of these specimens would indicate the nature of the processes of weathering, and also that it would be established whether unweathered crocidolite from a specific area has constant properties.

## PART 2. MINERALOGICAL INVESTIGATION. BY A. G. FREEMAN, A. HODGSON, AND H. F. W. TAYLOR

### PRELIMINARY INVESTIGATION

The samples were unbroken pieces weighing up to 4 kg each containing one or more seams of cross-fiber crocidolite between layers of massive or shaly dark blue rock. Individual seams were 3–40 mm thick. In the samples from the Fresh Zone (i.e. all except RS 1, 2, 4, 18 and 19) the rock was usually clean, and the fiber a very deep blue, with a distinct sheen. The samples from the Oxidized Zone (RS 1, 2, 4 and 19) and the Silicified Zone (RS 18) were visibly weathered in varying degrees, the rock displaying patches or streaks of iron oxides. In sample RS 18, the rock was much decomposed with formation also of nontronite and of siliceous and calcareous encrustations. The fiber seams appear much more resistant to weathering than the rock. The fiber from the Oxidized and Silicified Zones tended, however, to be dull or grayish blue, commonly contained streaks of iron oxides, and in extreme cases was completely replaced by olive green nontronite.

Crushed fiber specimens from most of the samples were examined optically and by X-ray single crystal methods. The results, which are described in the next section, confirmed that the fiber was crocidolite (i.e. fibrous riebeckite). Quartz and magnetite were nearly always present, both being distributed fairly uniformly among the fibers in amounts probably not exceeding 3 percent. Carbonates were also generally present, but were unevenly distributed. In three samples (RS 2, 3, and 19), a little goethite was also found among the fibers. Two of these three samples are from the Oxidized Zone.

The carbonate impurities occurred in places as isolated grains, and also as streaks of needles or thin flakes arranged parallel to the fibers. Microscopic examination of some of these crystals showed a cleavage pattern, twinning, and high birefringence similar to those of calcite. Chemical analysis of the carbonate impurity isolated from three samples of fiber

showed high molar ratios of $CaCO_3$ to $(MgCO_3 + FeCO_3)$, the values being 14.4 (RS 14), 9.3 (RS 16), and 5.3 (RS 23). For samples from the Fresh Zone, the total quantity of carbonate impurities associated closely with the fiber was usually around 1 percent, though extremes of 0.3 to 9 percent were found. These estimates were obtained from $CO_2$ analyses on 10 g

TABLE 3

X-RAY POWDER DATA ($CoK_\alpha = 1.790$ Å)

| Observed spacings (Å) and relative intensities | | Indices and calculated spacings (space group I 2/m) | |
|---|---|---|---|
| 9.3 | 3 | 020 | 8.98 |
| 8.42 | 10 | 110 | 8.41 |
| 4.90 | 3 | 011 | 4.87 |
| 4.52 | 5 | 040 | 4.49 |
| 3.88 | 3 | 031 | 3.86 |
| 3.67 | 2 | 121 | 3.65 |
| 3.43 | 6 | $23\bar{1}$ | 3.40 |
| 3.24 | 3 | 240 | 3.27 |
| 3.09 | 8 | 310 | 3.12 |
| 2.97 | 3 | 141 | 2.98 |
| | | $32\bar{1}$ | 2.96 |
| 2.79 | 2 | 330 | 2.80 |
| 2.719 | 10 | 231 | 2.71 |
| | | $25\bar{1}$ | |
| 2.602 | 4 | 161 | 2.59 |
| 2.530 | 6 | 002 | 2.53 |
| 2.327 | 4 | 251 | 2.32 |
| 2.261 | 4 | 112 | 2.26 |
| 2.173 | 5 | $36\bar{1}$ | 2.19 |
| | | $27\bar{1}$ | 2.18 |
| 2.132 | 1 | 132 | 2.13 |
| 2.077 | 2 | | |
| 2.012 | 2 | | |
| 1.990 | 1 | | |
| 1.892 | 1 | | |
| 1.868 | 2 | | |
| 1.816 | 1 | | |
| 1.799 | 2 | | |
| 1.726 | 1 | | |
| 1.697 | 1 | | |
| 1.683 | 1 | | |
| 1.655 | 5 | | |
| 1.637 | 1 | | |
| 1.612 | 2 | | |
| 1.591 | 4 | | |
| 1.575 | 1 | | |
| 1.518 | 3 | | |
| 1.503 | 2 | | |
| 1.493 | 1 | | |
| 1.459 | 1 | | |
| 1.424 | 5 | | |
| 1.377 | 3 | | |
| 1.349 | 4 | | |
| 1.309 | 5 | | |
| 1.285 | 2 | | |
| 1.239 | 1 | | |
| 1.202 | 1 | | |
| 1.192 | 1 | | |

portions of fiber made up of pieces taken from different parts of the hand specimen, all the impurities being included with the disintegrated fiber.

### *Optical and X-ray Examination*

Clean fibers from each of a representative range of samples (RS 1–8 and 18–21) were examined optically and by X-ray single crystal and powder methods. For the X-ray work, 6 and 11.46 cm cameras were used, with filtered cobalt radiation. Neither X-ray nor optical methods revealed any difference between the fibers from the various samples, although these included samples from Fresh, Oxidized, and Silicified Zones. The refractive indices of fiber bundles are 1.69 parallel to their *c*-axes, and 1.71 at right angles thereto. The bundles have negative elongation and parallel or near-parallel extinction. Pleochroism is strong: the fibers are indigo-blue to green-blue when parallel to the direction of vibration of the polarizer, and gray-blue to a delicate lilac color when perpendicular thereto. These results agree well with typical data for riebeckite.

TABLE 4

UNIT CELL DATA

| Axial convention and space group | Body-centered I 2/m | Face-centered C 2/m |
|---|---|---|
| a | 9.97 Å | 9.80 |
| b | 17.85 Å | 17.85 |
| c | 5.30 Å | 5.30 |
| a sin $\beta$ | 9.52 Å | 9.52 |
| $\beta$ | 107°30′ | 103°50′ |

Attempts to obtain true, single X-ray patterns were unsuccessful, even the thinnest usable fibers (5 $\mu$ thick) proving to be bundles with random orientation around the fiber direction. The *c*-axes of the individual crystallites were well aligned parallel to this direction. The fibers therefore gave good *c*-rotation photographs, but long exposures were needed by comparison with true single crystals of riebeckite of similar size in order to obtain patterns of a given intensity.

The possibility was considered that in some or all of the samples the crocidolite might be intimately intergrown in parallel growth with some other amphibole or mineral of related structure. The rotation photographs were therefore compared with standard ones for several likely impurities, as well as with ones of riebeckite. No mineral other than riebeckite could be detected.

The approximate unit cell parameters of riebeckite are established (2–4), and the powder and rotation data (Table 3) were therefore readily indexed and unit cell parameters for the present material obtained. Table 4 gives the results in terms of each of the axial conventions for amphiboles presently in common use (8-10). The powder data and cell parameters agree substantially with those reported for riebeckite or magnesio-riebeckite from other sources (2-7).

Table 3. Chemical Analyses.

| Sample | RS1 | RS2 | RS3 | RS5 | RS6 | RS6 | RS7 | RS7 | RS8 | RS13 | RS14 | RS16 | RS18 | RS21 | RS22 | RS23 | x | y |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Method | A | A | A | A | A | B | A | B | A | A | A | A | A | B | A | A | | |
| $SiO_2$ | 49.40 | 50.85 | 51.00 | 51.70 | 51.45 | 50.52 | 50.90 | 51.03 | 51.40 | 51.20 | 50.61 | 50.53 | 52.40 | 50.26 | 51.20 | 51.20 | 50.65 | 51.94 |
| $TiO_2$ | nd | nd | nd | nd | nd | 0.005 | nd | 0.004 | nd | nd | nd | nd | nd | 0.004 | nd | nd | nd | nd |
| $Al_2O_3$ | nil | nil | nil | nil | nil | nil | nil | nil | nil | nil | nil | nil | 0.15 | nil | nil | nil | nd | 0.20 |
| $Fe_2O_3$ | 19.57 | 18.15 | 17.80 | 17.70 | 17.52 | 16.07 | 16.85 | 16.72 | 17.43 | 17.56 | 17.56 | 16.74 | 19.05 | 17.49 | 15.96 | 16.45 | 18.55 | 18.64 |
| FeO | 18.25 | 19.55 | 20.40 | 20.30 | 19.60 | 19.79 | 20.50 | 21.05 | 19.60 | 19.95 | 19.70 | 19.90 | 17.05 | 19.94 | 19.75 | 21.36 | 17.85 | 19.39 |
| MgO | 1.60 | 1.80 | 1.00 | 1.46 | 1.08 | 2.59 | 1.06 | 2.19 | 2.05 | 2.21 | 2.59 | 2.38 | 2.00 | 2.34 | 3.34 | 1.49 | 2.30 | 1.37 |
| MnO | 0.12 | 0.06 | 0.07 | 0.05 | 0.06 | 0.11 | 0.05 | 0.11 | tr. | 0.07 | 0.06 | 0.04 | 0.04 | 0.12 | 0.04 | 0.07 | nd | nd |
| CaO | 1.60 | 1.20 | 0.30 | 0.60 | 1.32 | 2.72 | 1.45 | 0.63 | 0.78 | 1.11 | 1.10 | 1.62 | 0.59 | 1.30 | 0.65 | 0.76 | 0.75 | 0.19 |
| $Na_2O$ | 6.28 | 6.25 | 6.22 | 6.07 | 6.40 | 5.43 | 6.20 | 5.67 | 6.27 | 6.31 | 6.26 | 6.12 | 5.59 | 5.68 | 5.52 | 6.36 | 5.70 | 6.07 |
| $K_2O$ | 0.18 | 0.37 | 0.14 | 0.20 | 0.15 | nil | 0.20 | nil. | tr. | tr. | tr. | tr. | tr. | nil. | tr. | tr. | nd | 0.04 |
| $CO_2$ | 1.00 | 0.08 | 0.23 | 0.23 | 0.23 | nd | 0.20 | nd | 0.33 | 0.06 | 0.79 | 0.63 | 0.82 | nd | 0.84 | nil. | nd | nd |
| F | nd | nd | nd | nd | nd | nil. | nd | nil. | nd | nd | nd | nd | nd | nil. | nd | nd | nd | nd |
| $H_2O^{+105}$ | 2.75 | 2.26 | 2.51 | 2.18 | 2.19 | 2.36 | 2.37 | 2.40 | 2.37 | 2.20 | 2.28 | 2.26 | 2.26 | 2.39 | 2.65 | 2.98 | 3.70 | 2.58 |
| $H_2O^{-105}$ | 0.23 | 0.11 | 0.39 | 0.25 | 0.26 | 0.36 | 0.22 | 0.29 | 0.20 | 0.17 | 0.18 | 0.30 | 0.30 | 0.29 | 0.44 | nil. | 0.80 | 0.31 |
| Total | 100.98 | 100.69 | 100.06 | 100.74 | 100.28 | 99.96 | 100.10 | 100.09 | 100.43 | 100.77 | 101.03 | 100.52 | 100.61 | 99.81 | 100.42 | 100.62 | 100.30 | 100.73 |
| $Fe^{+2} : Fe^{+3}$ | 1.04 | 1.20 | 1.27 | 1.27 | 1.24 | 1.37 | 1.35 | 1.40 | 1.25 | 1.26 | 1.25 | 1.32 | 0.99 | 1.27 | 1.38 | 1.44 | 1.07 | 1.16 |

All analyses in this investigation represent the mean of two closely agreeing separate determinations.

Method A: Analyst, W. Benns, Cape Asbestos Company Ltd. Sample RS23 analyzed after extraction of carbonates in hot 10% acetic acid.

Method B: Analyst, W.H. Herdsman, Glasgow (FeO and $H_2O$ analyses by A.G. Freeman)

x : Crocidolite, Westerberg (Hall(1)). Analyst, Government Chem. Lab., Johannesburg.

y : Crocidolite, Kliphuis (Peacock (12)). Analyst, H.E. Vassar.

### *Chemical Analysis*

Table 5 gives analyses for fourteen samples. Crocidolite, when disintegrated, is light and fluffy, with a bulk density below 0.1 g/cc. It contains small amounts of magnetite, quartz, and carbonates, which tend to accumulate towards the bottom of a heap of disintegrated fiber. Excessive disintegration, or grinding to a powder, causes partial oxidation of the iron and may also affect the water content. In preparing material for analyses, two procedures were therefore adopted:

*Method A.*—This was generally used. Portions of the fiber seam or seams are broken into small pieces and then cut with scissors into short, coarse lengths, which are then let fall on to a piece of paper or 36-mesh sieve. The fiber is picked up by hand; this leaves a substantial proportion of the impurities behind. Optical examination suggested that material so purified contained about 99 percent of crocidolite. Excessive magnetite can also be removed with a hand magnet. In one case carbonates were removed by digestion with hot, 10 percent acetic acid, which seemed not to attack the crocidolite.

*Method B.*—This was used to obtain better purification of three samples. The material for the main part of the analysis (i.e. for everything except $H_2O$ and FeO, but including total iron) is ground to below 100 mesh in an agate mortar and then centrifuged successively in methylene iodide, to remove magnetite, and in a methylene iodide-bromoform mixture, to remove quartz and other light impurities. The purified material is washed in benzene followed by ether, and dried in vacuo at 70° C. For the FeO and $H_2O$ determinations, a separate portion was used consisting of fibers chosen individually under the microscope for minimum contamination with impurities. These were coarsely chopped with scissors and purified by centrifugation as above. The FeO was determined by using Cooke's method (11). This procedure avoided errors from grinding, but was too tedious to be usable for the main portion of the sample.

Comparison of the results for RS 6 and RS 7, for which both methods were used, shows only slight differences between the two methods. Method B gives slightly lower $Fe^{+3}$:$Fe^{+2}$ ratios, which can perhaps be attributed to incomplete removal of magnetite in Method A, together with slight oxidation of $Fe^{+2}$. The simpler Method A thus seems adequate, at least for comparative studies, though Method B is perhaps preferable where the results are to be used for calculation of cell contents.

### *Densities*

Three methods were used:

(i) *Weighing in Water.*—3–5 g pieces of unfiberized material were hand cleaned to present smooth surfaces free from loose fibers, and immersed in distilled water under vacuum until gas bubbles ceased to escape (2–3 hr). They were then weighed, immersed in the same air-free water.

TABLE 6

DENSITIES

| Specimen | Method | Density (g/cc) |
|---|---|---|
| RS 1 | (i) | 3.37 |
| 2 | (i) | 3.36 |
| 3 | (i) | 3.36 |
| 5 | (i) | 3.34 |
| 6 | (i) | 3.33 |
| 6 | (ii) | 3.29 |
| 6 | (iii) | 3.32 |
| 7 | (i) | 3.34 |
| 7 | (ii) | 3.34 |
| 7 | (iii) | 3.35 |
| 8 | (i) | 3.37 |
| 13 | (i) | 3.34 |
| 14 | (i) | 3.37 |
| 16 | (i) | 3.34 |
| 18 | (i) | 3.31 |
| 21 | (ii) | 3.34 |
| 21 | (iii) | 3.35 |
| 22 | (i) | 3.37 |
| 23 | (i) | 3.42 |

(ii) *Suspension in Clerici's Solution.*—Selected fibers about 1 mm long, free from quartz or magnetite were used. The liquids were boiled under reduced pressure to remove air and afterwards centrifuged.

(iii) *Gas Displacement.*—A constant pressure hydrogen displacement apparatus was used. 12–15 g portions of fiber were selected from the cleanest

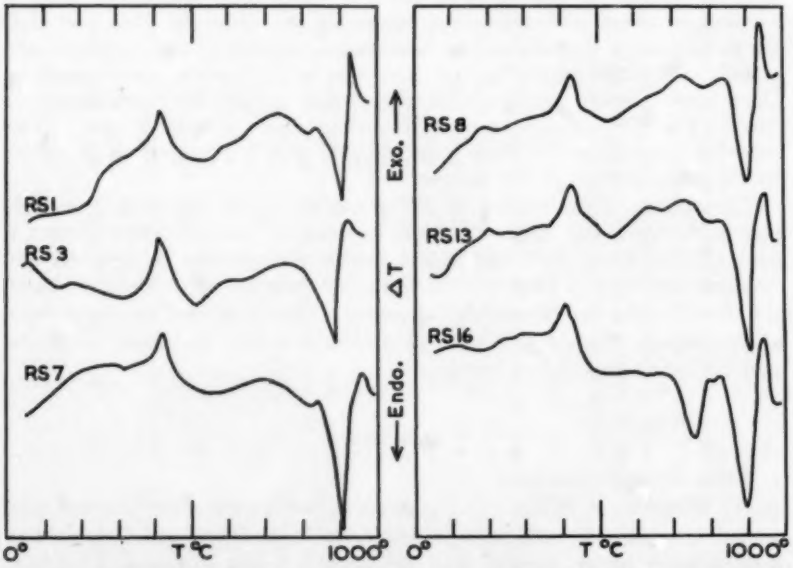


FIG. 3. Differential thermal analysis curves for specimens of fiber from Westerberg (left) and Koegas (right), obtained in air at 10 deg/min.

parts of the seams and cut with scissors into approximately 1 cm lengths. A trial of the method on calcite gave 2.71 g/cc (X-ray density 2.715 g/cc).

The results (Table 6) tend to fall into two groups around 3.34 and 3.37 g/cc, which can be correlated with the direction of fiber growth. The higher value is associated with growth perpendicular to the bedding planes of the rock, and the lower one with oblique growth, mostly at 45–70° to the bedding planes. The core sample (RS 23) had a significantly higher density than the others. No corrections were made for impurities with methods (1) and (iii). The errors due to quartz and magnetite will tend to cancel.

### *Differential Thermal Analysis*

DTA curves were determined for all of the samples in air at 10 deg/min. Typical curves (Fig. 3) show an exotherm at 410–420° C, a sharp endotherm at 895–910° C, and a small exotherm at 920–960° C. The endotherm at 750–820 C°, which sometimes occurs, is caused by carbonate impurities, the decomposition temperature of which falls as the proportion of siderite in them increases.

## DISCUSSION

### *Uniformity of the Material*

The samples examined represent a vast amount of workable ore, and the results show that there is little variation in the composition of crocidolite from the Outer, Main, and Inner Reefs in the Koegas-Westerberg area. The optical properties, densities and cell parameters likewise show little or no variation. The mineral from the Bottom Visser Reef (Hounslow Mine) is distinctive in its lower $Na_2O$ and higher MgO contents, but the difference is slight.

There is a rough correlation between the depth of occurrence and the $Fe^{+2}$:$Fe^{+3}$ ratio (Table 5), which is highest in sample RS 23, from a deep borehole, and lowest in samples RS 1, 2, and 18, from the Oxidized and Silicified Zones. This suggests that percolating ground waters may have caused slight oxidation, even at a depth of 300 feet (100 m). The three samples from the Weathered Zones also showed slightly more variation in $SiO_2$ content. Even these samples, however, differed little in composition from the others.

The most recent previous analyses for crocidolite from Cape Province are the groups reported by Hall (1) and Peacock (12). Typical analyses from each group, for samples from the Koegas or neighboring areas are included in Table 5. Both show more $Fe_2O_3$ and less FeO than reported here, and Peacock's analysis also shows more $SiO_2$. This suggests that these earlier analyses relate to partly weathered material.

### *Unit Cell Contents and Parameters*

The calculation of cell contents is complicated by several factors. The observed water contents ($H_2O^{+105}$) exceed the theoretical values (about

1.94%), and it is uncertain how this extra water is held. Even the freshest samples studied may have suffered slight oxidation, and this further complicates the interpretation of the water contents. Lastly, the observed densities are likely to be low for such fibrous material. The first two of these factors are being studied, and for the present only provisional calculations of cell contents and X-ray densities can be made. It will be assumed that the unit cell contains 48 oxygen and 4 hydrogen atoms, excess water being present on surfaces or in pockets outside the crystal structure. Table 7 gives the results for each of the three samples analyzed by Method B. The chemical composition approximates the $Na_2Fe'''_2Fe''_{2.5}Mg_{0.5}Si_8O_{22}(OH)_2$; i.e. riebeckite with one sixth of the ferrous iron replaced by magnesium.

Table 7. Atomic cell contents.

| | RS 6 | | | RS 7 | | | RS 21 | | |
|---|---|---|---|---|---|---|---|---|---|
| Si | 15.70 | | 16.00 | 15.84 | | 16.00 | 15.68 | | 16.00 |
| Fe* | 3.76 | 0.30 | | 3.91 | 0.16 | | 4.12 | 0.32 | |
| Fe* | 5.13 | 3.46 | | 5.48 | 3.75 | | 5.18 | 3.80 | |
| Ti | 0.01 | | 10.00 | 0.01 | | 10.00 | 0.01 | | 10.00 |
| Mn | 0.04 | | | 0.04 | | | 0.04 | | |
| Mg | 1.22 | | | 1.01 | 0.72 | | 1.08 | 0.97 | |
| Ca | 0.91 | 0.14 | | 0.20 | 0.29 | | 0.43 | 0.11 | |
| Na | 3.28 | 0.77 | 4.05 | 3.41 | | 3.90 | 3.42 | | 3.96 |
| H | 4.00 | | | 4.00 | | | 4.00 | | |
| O | 48.00 | | | 48.00 | | | 48.00 | | |
| X-ray density g/cc | 3.43 | | | 3.43 | | | 3.43 | | |

There is no evidence as to whether the distribution of cations among the octahedral sites is ordered; in other monoclinic amphiboles studied recently no such ordering has been found (13, 14).

Whittaker (8) showed that linear relations exist between $\beta$ and a $\sin\beta$, and between $\beta$ and the radius of the cations in the X-sites (Na in riebeckite), for those monoclinic amphiboles for which data were available. The present results agree well with his theory for the second of these relations, and moderately well for the first.

### *Thermal Behavior*

The DTA curves (Fig. 3) resemble those previously reported for crocidolite (7, 15). Vermaas (7) attributed the 410–420° C exotherm to partial oxidation of the ferrous iron, and the large endotherm at 895–910° C to dehydration. He attributed the 920–960° C exotherm to crystallization of anyhydrous products, possibly accompanied by oxidation of the remaining ferrous iron. We are making further studies to test this interpretation.

These suggest, in accordance with the interpretation of the chemical analysis already given, that part of the water retained above 105° C is uncombined, as its loss does not seem to affect the crystal structure. Loss of this water, at temperatures up to 400° C, whether in air or in inert atmospheres, does however cause progressive diminution in fiber strength.

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# VOLUME RELATIONSHIPS DURING REPLACEMENT REACTIONS[1]

L. L. AMES, JR.

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ABSTRACT

The so-called "Law of Equal Volumes" is a result of differences in bulk density and molar volume relationships between initial and final replacement reaction products. The replacement of calcite by whewellite, for example, proceeds in a kinetic system until the large positive volume change from calcite to whewellite begins to affect the permeability of the original calcite body. Eventually the flow of oxalate-containing solution may cease entirely and the resulting "whewellite replacement body" apparently represents volume replacement of calcite by whewellite.

Replacement reactions and initial rates are determined by the activities of the system constituents in relation to their activity product ratios. Large positive or negative volume changes between initial and final products do not initially prevent or cause replacement reactions, but are a result of replacement reactions.

## INTRODUCTION

THE possibility of a volume reduction from initial mineral to replacement product was recognized by Irving (8) who listed "volume vugs" as a criterion of replacement. Several authors since have reported volume reductions as a result of replacement reactions. Currier (4) noted and explained a volume reduction in the Illinois fluorspar field as probably due to stoichiometric replacement of limestone by fluorite. Grogan (7) explained some of the structural features of the same fluorspar field as being a result of stoichiometric limestone replacement. Gillerman (6) also suggested a volume reduction observed in the Eagle Mountain fluorspar deposits as due to stoichiometric limestone replacement by fluorspar. Schouten (12), on the other hand, reported that volume losses due to replacements of up to 30 per cent often could not be detected under the microscope. Enterolithic folding, especially during alteration of anhydrite to gypsum, is not uncommon according to Petti-

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john (11). Garrels and Dreyer (5) reported that due to a large positive molar volume change in going from calcite to cerrusite, they obtained no reaction between calcite and $PbCl_2$ or $Pb(NO_3)_2$ solutions. These authors postulated that a monomolecular layer of lead carbonate was formed on the surface of the calcite preventing further reaction between calcite and solution. Volume relationships between minerals involved in replacement reactions have been summarized in "The Law of Equal Volumes" which states that the replacing mineral occupies exactly the space formerly filled by the primary mineral (10).

During the study of several metasomatic replacement reactions (1, 2, 3), thin-sections were prepared showing the alteration of an original mineral to some more stable alteration product. Examination of these thin-sections revealed that interpretations of replacement reaction kinetics based on the "Law of Equal Volumes" often could not be substantiated by the experimental evidence. Consequently the relationship of volume of initial mineral to alteration product, and possible effects of this relationship on the reaction kinetics was investigated. The results and conclusions of this study are presented in this paper.

## METHODS OF INVESTIGATION

The method of investigation used in this study was reported previously along with several of the reactions (2). The reactive salt in solution, traced with $Sr^{85}$, $Ba^{133}$, $Zn^{65}$ or $C^{14}$ as appropriate, was passed through a mineral column by means of a constant-head pump. The tracing cations were present as chlorides while the $C^{14}$ was added as $Na_2C_2{}^{14}O_4$. A 100 ml volume of the column effluent solution was collected and analyzed for the radioactive constituent.

The difference between the concentration of radioisotope that went into the top of the column and that collected after passage through the column was assumed to represent the fraction of original radioisotope concentration reacting with the column material. The reactive constituent of the solution was accompanied by a mole/liter of NaCl to minimize side reactions such as ion exchange or adsorption. Activity products were calculated from $\Delta F°$ values given by Latimer (9).

Molar volumes were calculated from specific gravities. For example, the molar volume of calcite, 36.93 cc/mole, is calculated as follows:

$$\frac{(100.09 \text{ g/mole})}{(2.71 \text{ g/cc})} = 36.93 \text{ cc/mole}.$$

When reactions are simple mole-for-mole substitutions, as they are in the reactions discussed here, volume changes can be computed directly from the molar volumes of the minerals involved.

## RESULTS

Figure 1 shows the replacement of a single crystal of calcite by fluorite. Note that the single crystal of calcite is being replaced by many smaller fluorite crystals, and there are spaces between the elongated fluorite crystals

through which fluoride and carbonate ions moved. These crystals obviously could not have grown in a direction ninety degrees to their present elongate axes without shutting off their own growth nutrient, a circumstance that would account for the unusual crystal habit. Figure 1 also shows that the bulk density of the placing fluorite approaches the theoretical density of the single calcite crystal.

Figure 2 is a photomicrograph of the replacement of a single calcite crystal by whewellite. The original physical character of the calcite is nowhere revealed by the replacing whewellite; i.e., all original physical characteristics of the calcite were destroyed during the replacement process—unlike a previously reported (3) fluorite replacement of calcite where even the interfacial angles of calcite were faithfully preserved.

Figure 3 shows the effect of temperature on some cation and anion replacement reactions. Because a common unit of time is involved, Figure 3 also represents comparative reaction rates for the various reactions. Only replacement reactions having an anion or cation common to both initial and final

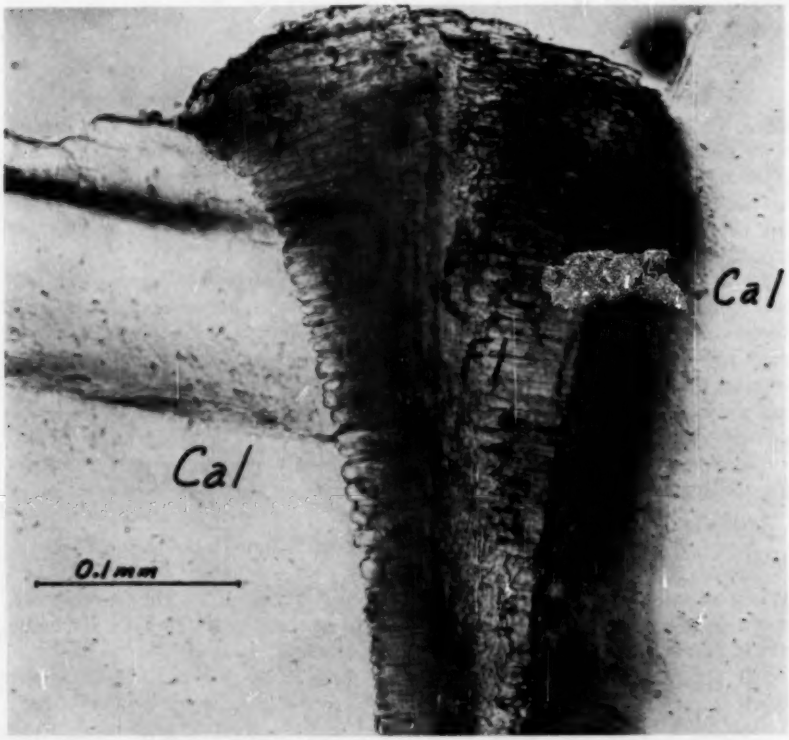


FIG. 1. Photomicrograph of the replacement of calcite by fluorite. Cal = Calcite, Fl = Fluorite.

FIG. 2. Photomicrograph of the placement of calcite by whewhellite. Cal = Calcite, Wh = Whewellite.

minerals are discussed here. X-ray diffraction patterns of the end-product of the kinetic reactions between calcite and $Pb(NO_3)_2$-containing solutions demonstrated that the reaction readily proceeds at room temperature to form $PbCO_3$. Lack of a usable Pb radioisotope discouraged detailed study of the reaction.

The following represent the reactions shown in Figure 3:

1) $\underline{CaSO_4 \cdot 2H_2O} + BaCl_2 \rightleftharpoons \underline{BaSO_4} + CaCl_2 + 2H_2O$

2) $\underline{CaCO_3} + ZnCl_2 \rightleftharpoons \underline{ZnCO_3} + CaCl_2$

3) $\underline{CaCO_3} + Na_2C_2O_4 \rightleftharpoons \underline{CaC_2O_4} + Na_2CO_3$

4) $\underline{CaCO_3} + SrCl_2 \rightleftharpoons \underline{SrCO_3} + CaCl_2$

5) $\underline{CaCO_3} + BaCl_2 \rightleftharpoons \underline{BaCO_3} + CaCl_2$

Solid minerals retained in the column are underlined.

DISCUSSION

The author has stated in previous papers (1, 2) that the order and rate of replacement are a function of activity product ratios between initial and final replacing mineral constituents in solution and the activities of their respective constituents in solution. Other authors (10) have contended that the so-called "Law of Equal Volumes" expresses the basic relationships involved in replacement processes, implying that such processes are non-stoichiometric.

Evidence available from thin-section examinations of replaced minerals both support and disclaim the "Law of Equal Volumes." Figure 1 satisfies the "Law of Equal Volumes" even though there is a volume change of − 30 percent during the reaction, because the bulk density of the replacing fluorite approaches the theoretical density of the original calcite. The whewellite re-

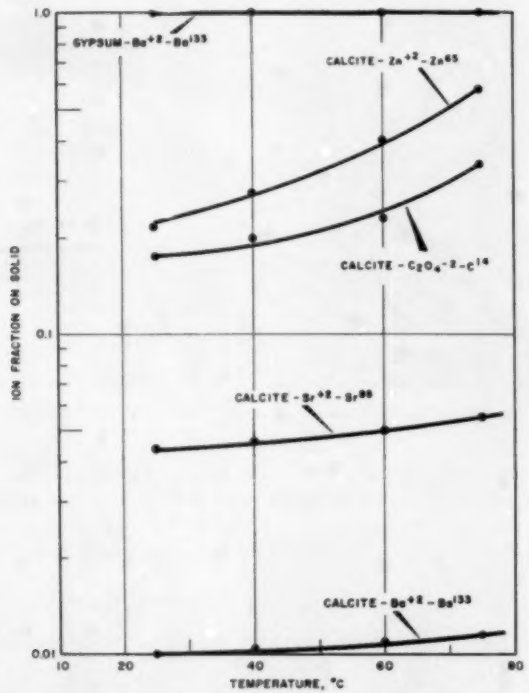


FIG. 3. A comparison of the initial rates and order of replacement of several replacement reactions. Influent solution—1 *M* NaCl, 0.05 *M* other active macro-constituents, radiotracers as indicated; 100 ml samples collected. Influent pH—5.0. Common flow rate—410 ml/cm²/hr. Temperature—as indicated. Initial column—50 g, 1.0 to 0.25 mm calcite or gypsum as indicated, 2.84 cm² cross-sectional area. Surface areas—0.067 m²/g.

TABLE I

| Reaction<br>Molar volumes $V_1$ $V_2$<br>Activity products (9) $a_1$ $a_2$ | Percent loss or gain in volume from $V_1$ to $V_2$ | $\text{Log}_{10}(a_1/a_2)$ | Initial ion fraction reacted at 25° C (from Fig. 3) |
|---|---|---|---|
| Gypsum → Barite | −30.2 | +4.38 | 0.999 |
| Calcite → Smithsonite | −23.6 | +1.36 | 0.223 |
| Calcite → Whewellite | +80.0 | +1.08 | 0.182 |
| Calcite → Strontianite | + 7.9 | +0.83 | 0.044 |
| Calcite → Witherite | +20.8 | +0.47 | 0.012 |

placement of Figure 2 apparently does not satisfy a volume-for-volume relationship. The large positive change in volume from a single calcite crystal to whewellite discredits the concept of equal volume replacements. Field observations of replacements may be noted that also discredit or substantiate the "Law of Equal Volumes." To resolve the problems of equal volume replacements, a closer examination of volume changes and activity product ratios and their relationship to the order and rate of the replacement reactions shown in Figure 3 is in order.

Table I presents the volume changes and activity ratios applicable to the reactions listed in Figure 3. The correlation coefficient between volume change and initial fraction of cation reacted is − 0.51, while the correlation coefficient between log activity product ratios and fraction of cation reacted is + 0.99. Volume relationships are a result of a given replacement reaction, but do not initially cause or prevent the reaction.

Volume changes do, however, speed or retard an ongoing replacement reaction rate by changing the permeability of the altering mineral. For example, 0.1 mm calcite reacts with $Na_2C_2O_4$-containing solution eventually to "cement" the altering calcite gains, due to the large volume expansion, until the pump can no longer circulate the altering solution through, the column. As long as space is available for the oxalate solution to circulate, the alteration of calcite to whewellite continues through the whewellite alteration layer despite the large volume expansion. The apparent result of the whewellite reaction is a volume-for-volume replacement of much of the calcite.

In other words, as the replacement reactions in Table I proceed, the correlation coefficient between fraction of ion reacted and per cent loss or gain in volume begins to approach − 1.0, due to the effect of molar volume changes on reaction kinetics.

Starting with equivalent experimental conditions, such as surface area, the initial significant correlation is between ion fraction reacted and activity product ratios. As the reactions proceed, the significant correlation tends to be between ion fraction reacted and molar volume change.

There are at least two explanations for, or causes of, the "Law of Equal Volumes": 1) replacement reactions exhibiting negative theoretical volume

changes often show little, if any, actual volume change even at atmospheric pressure due to bulk density relationships between the original and replacing mineral, and 2) reactions involving theoretical positive volume changes decrease the permeability of the original mineral, forcing the replacement to become diffusion-dependent, and hence very slow. The larger-volume, replacing mineral, due to reaction kinetics, tends to occupy the same actual volume as the original mineral.

Unless the bulk density of both the original and replacing minerals is known, volume relationships are an unknown quantity. To illustrate the importance of bulk densities to replacement reaction volume relationships, let us compute the volume relations in the hypothetical replacement of a pure limestone bed by andradite garnet of the composition $3CaO \cdot Fe_2O_3 \cdot 3SiO_2$. Theoretical densities will be assumed for the limestone and garnet. The calcite-to-garnet molar ratio is 3 calcite $\rightarrow$ 1 garnet. The volume occupied by the original limestone is

$$\frac{100.09 \text{ g/mole}}{2.79 \text{ g/cc}} = 36.93 \text{ cc/mole},$$

$$(36.93 \text{ cc/mole})\ (3 \text{ moles}) = 110.8 \text{ cc calcite volume}.$$

The volume occupied by the resulting andradite, utilizing all available CaO, is

$$\frac{508.11 \text{ g/mole}}{3.75 \text{ g/cc}} = 135.5 \text{ cc/mole},$$

$$(135.5 \text{ cc/mole})\ (1 \text{ mole}) = 135.5 \text{ cc garnet volume}.$$

The assumption is made that the $Fe_2O_3$ and $SiO_2$ were added from an outside source and that the remaining $CO_3^{-2}$ was removed from the reacting limestone area.

The volume change from calcite to garnet is a positive one,

$$\begin{array}{r} 135.5 \text{ cc garnet volume} \\ -110.8 \text{ cc limestone volume} \\ \hline + 24.7 \text{ cc volume change, or } + 22.3 \text{ percent.} \end{array}$$

If we assume that a volume-for-volume replacement has occurred, and the volume of the original limestone was also 135.5 cc, then the specific gravity of the original limestone was 2.22 and not 2.71. Volume-for-volume replacement could be quite feasible in this case with a reasonably small deviation from theoretical limestone density.

It is unnecessary to assume that any known chemical principles are being violated or superseded during replacement reactions. While the "Law of Equal Volumes" often expresses volume relationships during replacement, those reactions leading to negative volume changes can result in an overall volume shrinkage.

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RICHLAND, WASHINGTON,
*July 1, 1961*

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# TEMPERATURES OF FORMATION AND ORIGIN OF THE NIGADOO AND BRUNSWICK MINING AND SMELTING NO. 6 DEPOSITS, NEW BRUNSWICK, CANADA[1]

J. KALLIOKOSKI

ABSTRACT

The Nigadoo deposit is a telescoped vein at and near the margin of an epizonal quartz-feldspar porphyry body. The vein minerals are coarse-grained and consist mainly of pyrrhotite, pyrite, sphalerite, galena, chalcopyrite, arsenopyrite, and stannite in a multi-generation calcite gangue. The iron content of sphalerite suggests a temperature of formation of 670° C, which contrasts with the obvious low-metamorphic rank of the wall-rocks.

The B. M. and S. No. 6 deposit, consisting essentially of pyrite with galena and sphalerite, is a large, lenticular, massive, fine-grained replacement in greenschist wall-rocks. Along the footwall an adjoining zone contains pyrrhotite, chalcopyrite, and a little sphalerite and galena. In general the iron content of the sphalerite is higher where it occurs with pyrrhotite than where it accompanies pyrite, but details of spatial and compositional variation suggest that the relationships are considerably more complex. Sphalerite compositions suggest a maximum temperature of formation near 480° C which would exceed the suggested upper limit of the greenschist facies by about 250° C.

The writer believes that the temperature data proves that both deposits are of epigenetic origin.

## INTRODUCTION

THIS paper presents some results in trying to apply the sphalerite geothermometer (13) to certain New Brunswick mineral deposits. The work was undertaken in order to gain some insight into problems connected with geothermometry. It developed a specific problem concerning the significance of temperatures, which in the present case appear to be too high. This study forms part of a much larger research project undertaken by the Geological Survey of Canada under the direction of R. W. Boyle, the purpose of which is to study the geology and geochemistry of New Brunswick mineral deposits.

The sphalerites were collected by the writer from the Nigadoo Mine and from the Brunswick Mining and Smelting No. 6 deposit.

The writer wishes to express his appreciation to the Geological Survey of Canada for employing him for the field-mapping and collecting phase of this project. The writer would also like to thank Messrs. Francis Cameron, Norman H. Donald, Jr., and E. R. Lea of the St. Joseph Lead Company, Dr. J. J. Aleva then of Nigadoo Mines, and Dr. D. R. E. Whitmore, then of Heath

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Steele, for help of various kinds. Natural, analyzed sphalerites were supplied to the writer by A. A. Chodos and S. Berman, and synthetic ones by P. Barton. Other analyses, by Lucius Pitkin, Inc., were financed by research funds of the Department of Geology, Princeton University. Throughout this work the writer has received much help and encouragement from H. D. Holland, and the paper has benefitted from critical reading and editing by E. Sampson and R. W. Boyle.

In the field the writer was helped by R. Crockett and in the laboratory at various times by R. L. Chase, M. S. Joshi, and S. B. Lumbers. To all he expresses his sincere thanks.

The analytical procedure finally adopted by the writer is similar to that of Chodos, Rose (16), and Benson (6, 7). About one to three milligrams of sphalerite were dug out with a sharp needle, under a microscope, and mounted on ½ mil mylar film with a drop of acetone-diluted rubber cement. The intensity of FeK$\alpha$ and second order ZnK$\alpha$ was compared with a working curve obtained from chemically analyzed sphalerites. Using 3 mg samples of sphalerite it was found with repeated X-ray fluorescence analyses of a particular sample that the variation in iron content determinations was equivalent to about $\pm 12°$ C.

The manganese and cadmium contents of sphalerite was determined by filling a small plastic holder with the specimen, and measuring the intensity of the respective lines. The intensities were compared with two working curves, prepared by admixing known quantities of chemically pure MnS and CdS with ZnS.

Because of X-ray tube contamination, copper had to be determined by wet chemical means.

## NIGADOO MINE

The Nigadoo deposit is a vein-structure, averaging about 3 feet in width and following a fault along and near the contact between an unsheared, epizonal quartz-feldspar porphyry body, and steeply dipping argillites and slates of Silurian age that show very little contact-metamorphic effects. Along the vein the porphyry is sericitized and kaolinized for short distance into the walls.

The vein is variable in character both vertically and horizontally. In places it consists of massive sulfides, commonly without inclusions of wall-rocks, but locally it exhibits a well-developed layering, due in part to sequential deposition and in part to slight shearing. Locally the sulfides extend into the wall-rocks along tiny cross-cutting fractures. A segment of the vein, consisting of zoned, coarsely crystalline calcite, shows a well-developed cockade structure with several generations of brecciation and recementation, the calcite both cutting and being cut by sulfides. According to Aleva (2) pyrrhotite is the predominant iron sufide at depth whereas pyrite with minor pyrrhotite is characteristic of the upper levels (Fig. 1).

The sulfides are very coarse grained. The paragenetic sequence as determed by Aleva [2] and Roy [3] is shown in Table 1, but the writer is not cer-

[2] J. J. Aleva, personal communication.
[3] S. Roy, personal communication.

tain which of these is to be preferred. The second is more in harmony with some observations made by the writer, with the difference that the writer agrees with Aelva that arsenopyrite is younger than both pyrrhotite and sphalerite.

The minerals are appropriate for the hypothermal class but the rebrecciated vein structure is diagnostic of rather shallow depth of formation. Hence the deposit might be described as being "telescoped," showing both high tempera-

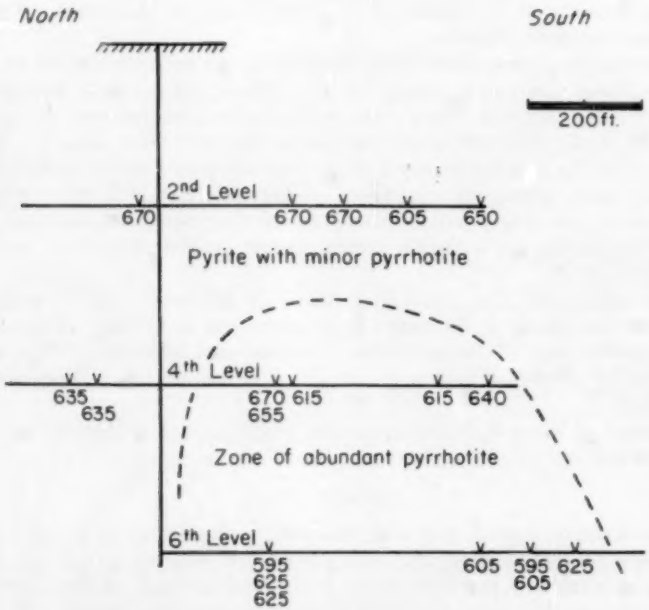



Figure 1. Longitudinal section of Nigadoo Mine showing sphalerite temperatures (°C).

Boundary of the pyrrhotite zone is from Aleva.

ture (hypothermal) and shallow (meso- to epithermal) characteristics. Because of the lack of distinctive textures, the vein is not representative of the xenothermal class (8) as the writer has suggested elsewhere (12).

The sphalerite analyses are listed in Table 2. As indicated, pyrite, pyrrhotite, and sphalerite coexist in many specimens, but paragenetic relationships do not indicate simultaneous deposition.

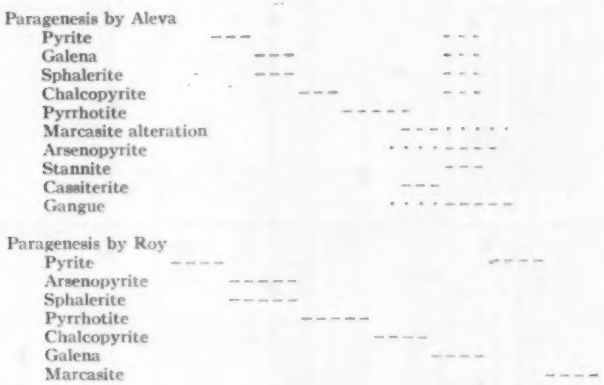
The sphalerite contains exsolved chalcopyrite, which shows a characteristic uneven distribution within a single grain, and the quantity of which varies considerably from one specimen to another. The exsolved chalcopyrite occurs

as lamellae, and as microscopic to submicroscopic blebs that produce a clouding effect. Exsolved pyrrhotite occurs as sporadic ovoid blebs and stannite is present as well-oriented blebs and lamellae.

The material analysed for iron was obtained from areas that showed no chalcopyrite clouding but which may have contained exsolved pyrrhotite and chalcopyrite. An attempt was made to avoid areas containing appreciable exsolved stannite. The sphalerites submitted for copper analysis were almost free of exsolved chalcopyrite and were chosen in order to find out how much chalcopyrite might have been retained in solid solution.

Four of the sphalerites were checked for the presence of wurtzite by X-ray diffraction, but none of the lines were present (19). Two of these sphalerites

TABLE 1




have a d-spacing of 5.426 Å, corresponding to more than 36 mole per cent FeS (18). If a correction is made for the cadmium and manganese content of the sphalerite these values become 5.424 Å and 32 mole per cent FeS respectively.

Nine sphalerites were analyzed for manganese and cadmium and four for copper.

In the hope of obtaining confirmatory temperature data, X-ray diffraction patterns were obtained of four pyrrhotites (3, 4). Unfortunately the pyrrhotites had inverted to a possibly monoclinic form and hence were unsuited for this purpose.

*Evaluation of Results.* Two features stand out: the generally high temperatures indicated by the iron content of the sphalerites, and the suggestion in the data (Fig. 1) that higher temperatures occur nearer the surface. How reliable are the data?

The quantities of Cd, Mn, and Cu in sphalerite are so low that these should not affect the solubility of FeS in sphalerite. Thus according to the work of Barton and Kullerud (5), the iron compositions represent real, minimum temperature values. However, some of the sphalerite also shows exsolution

bodies of stannite, and at present there is no information available regarding the effect of Sn on the solubility of FeS. Perhaps it is not serious.

Clark has shown (9) that arsenopyrite and pyrite cannot coexist above 491° C. If the original temperature was about 670° C as here suggested, then the temperature must have dropped about 180° C by the time that arsenopyrite was crystallizing. Temperatures could not have been above 491° at any time after arsenopyrite had formed.

Table 2

Sphalerite analyses, Nigadoo Mine

| Location | Minerals with sphalerite | | | | | | | Fek / Znk(2) | Mol % FeS | T°C | Mol % CdS | Mol % MnS | Wt.% Cu | Diffraction | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Pyrite | Pyrrhotite | Arsenopyrite | Chalcopyrite | Galena | Marcasite | Stannite | | | | | | | sl. vs wurts. | d | Mol % FeS |
| 27' S 325 | x | | x | x | | | v | 1.79±.05 | 24.6 | 670 | .85 | | 0.11 | sl | 5.426 | 36.5 |
| E wall 327 | x | | | v | x | x | | 1.77 | 24.6 | 670 | .85 | .67 | | | | |
| 30' N 329 | x | v | x | v | x | x | | 1.50±.04 | 21.2 | 605 | .85 | .67 | | | | |
| 303 S | x | | x | x | | | | 1.81 | 24.6 | 670 | .85 | .67 | | | | |
| S of 333 | x | v | x | v | x | x | v | 1.70 | 23.1 | 650 | | | | | | |
| 62' N 606 | x | | x | | | | | 1.65 | 22.7 | 635 | .65 | | | | | |
| 10' S 619 | x | v | x | x | | | v | 1.55 | 21.6 | 615 | .55 | .56 | 0.45 | sl | | |
| 605 | x | v | x | v | x | | | 1.80 | 24.6 | 670 | | | | | | |
| | | | | | | | | 1.73 | 23.2 | 655 | | | | | | |
| 608 | x | | x | x | | | | 1.64 | 22.6 | 635 | | | | | | |
| 35' S 621 | x | x | x | x | x | | v | 1.66 | 22.8 | 640 | | | | | | |
| 10' S 605 | x | x | x | x | x | x | v | 1.54 | 21.6 | 615 | | | | | | |
| 917 | x | x | x | v | x | x | | 1.51 | 21.2 | 605 | .85 | .56 | | | | |
| 921 | x | x | x | v | x | | x | 1.46 | 20.7 | 595 | .70 | 1.2 | 0.20 | sl | 5.4264 | 37.5 |
| | | | | | | | | 1.52 | 21.2 | 605 | | | | | | |
| 42 S 903 | x | x | | x | x | x | v | 1.46±.04 | 20.7 | 595 | .55 | .67 | 0.17 | sl | | |
| | | | | | | | | 1.58 | 22.0 | 625 | | | | | | |
| | | | | | | | | 1.58 | 22.0 | 625 | | | | | | |
| 31' S 923 | x | x | x | x | | x | | 1.59 | 22.0 | 625 | | | | | | |

x = major component
v = exsolution phase

Although the accompanying data suggest a slight increase in temperature toward the surface there is not enough evidence to prove this. This apparent increase may not be significant.

With the exceptions that the effect of Sn on the system is not known, and no correction has been made for depth of formation, it appears from this work that the Nigadoo deposit was formed at 670° C, a temperature which the writer accepts with considerable reluctance. The source for this deposit cannot be very far away.

### B. M. AND S. NO. 6 DEPOSIT

This deposit is about 1,100 feet long and 400 feet wide at the surface, at least 1,000 feet deep, and contains a minimum of 28,000,000 tons of mixed, massive sulfides (14). The deposit consists of two parts: one, situated along a southwest plunging anticlinal nose, is made up essentially of pyrite, sphalerite,, and galena; the other in direct contact with the former, and lying along the footwall shear-zone, consists predominantly of pyrrhotite and pyrite, with

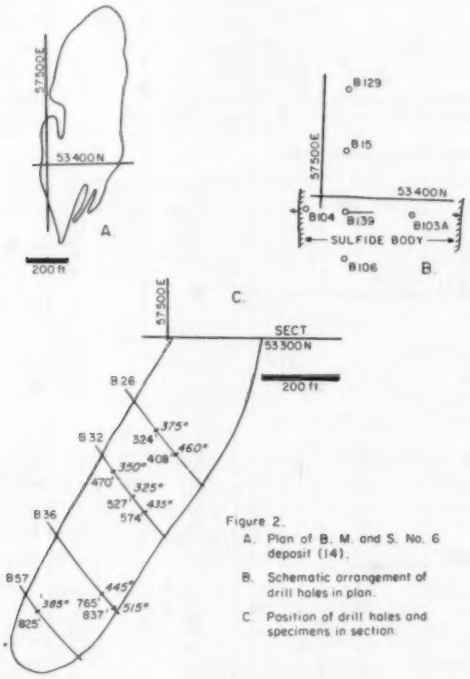



Figure 2.
A. Plan of B. M. and S. No. 6 deposit (14).
B. Schematic arrangement of drill holes in plan.
C. Position of drill holes and specimens in section.

lesser amounts of chalcopyrite, sphalerite, and galena. The exact temporal and structural relationships between these two bodies and assemblages are not clear (15, 21).

There is still a third type of mineralization at the No. 6 deposit which is mentioned only briefly by Stewart (21). This is a breccia ore consisting of brecciated pyrite that is healed by zoned veinlets of coarse galena, chalcopyrite, and pyrrhotite, galena being the earliest of the three. This type is rare and none of it was seen by the writer.

In all, the following 21 opaque minerals have been reported from this deposit: pyrite, sphalerite, galena, pyrrhotite, chalcopyrite, and magnetite, with

lesser amounts of stannite, cassiterite, tennantite-tetrahedrite, boulangerite, enargite, native silver, cubanite, bornite, chalcocite, covellite, gold, arsenopyrite, specular hematite, bismuth and bismuthinite (1, 14, 20).

The samples used in the present study were collected from drill-cores that cut the body, with the view of trying to determine horizontal and vertical

Table 3

Sphalerite analyses, B. M. and S. No. 6 deposit

| Drill hole | Footage | Minerals with sphalerite: Pyrite | Galena | Magnetite | Pyrrhotite | Chalcopyrite | Stannite | FeK$_\alpha$ / ZnK$_\alpha$(2) | Mol % FeS | T°C | Benson (6) |
|---|---|---|---|---|---|---|---|---|---|---|---|
| **Near surface, across strike** | | | | | | | | | | | |
| B-104 | 53 | X | X | | | | | .589;.617 | 9.8 | 335 | |
| | | | | | | | | .625 | 10.3 | 350 | |
| | | | | | | | | .637 | 10.5 | 355 | |
| | 54 | X | X | | | | | .626 | 10.3 | 350 | |
| | 241 | X | X | | v | | x | .671 | 11.0 | 370 | |
| B-139 | 21 | X | X | | | | | .74 ±.02 | 12.0 | 395 | |
| B-103A | 19 | X | X | | | | | .772 | 12.5 | 410 | |
| **Along strike, center-line of body** | | | | | | | | | | | |
| B-106 | 264 | X | X | | | | | .707 | 11.6 | 385 | |
| | | | | | | | | .678 | 11.2 | 375 | |
| B-139 | | | | | | | | | | 395 | |
| B-15 | 32 | X | X | | | | | .433 | 7.3 | 255 | 310° |
| B-129 | 61 | X | X | | | | | .675 | 11.0 | 370 | |
| | 93 | | | | | | | .842 | 13.5 | 435 | |
| B-57 | | | | | | | | | | | |
| **Down dip, center-line of body** | | | | | | | | | | | |
| B-26 | 324 | X | X | | | | | .678 | 11.1 | 375 | 385° |
| | 408 | X | X | | | X | | .928 | 14.6 | 460 | 450°-460° |
| B-32 | 470 | X | | | | | | .625 | 10.3 | 350 | 380° |
| | 527 | X | X | | | | | .571 | 9.5 | 325 | |
| | 574 | X | X | | | | | .865 | 13.6 | 435 | |
| B-36 | 765 | X | | | X | | | .885 | 14.0 | 445 | 700'-295 |
| | 837 | X | | X | X | X | | 1.12±.04 | 17.0 | 515 | |
| B-57 | 823 | | | | | | | .702 | 11.6 | 385 | |
| **Additional samples** | | | | | | | | | | | |
| B-8 | 146 | X | X | v | v | | | 1.04±.06 | 15.6 | 490 | |
| B-11 | 128 | X | X | v | v | | | .93±.06 | 14.6 | 460 | |
| B-27 | 627 | X | X | | v | X | | .972 | 15.3 | 480 | |
| B-30 | 419 | X | X | | v | v | | .674 | 11.1 | 375 | |
| B-33 | 676 | X | X | X | | | | .940 | 14.8 | 460 | |
| B-37 | 706 | X | X | | | | | .583 | 9.8 | 335 | 725'-340° |
| B-38 | 663 | X | | | X | | | .880 | 14.0 | 445 | 500'-450° |
| | | | | | | | | | | | 550'-495° |
| | 977 | | | | | | | .710 | 11.6 | 385 | 500'-450 |
| B-47 | 303 | | | | | | | .884 | 14.0 | 445 | |

X = major component
v = exsolution phase

temperature zonation by means of measurable changes in the iron content of the sphalerites (Fig. 2). Because of the low copper content of the body, most of the sphalerites examined were almost lacking in exsolved chalcopyrite, but where sphalerite is associated with pyrrhotite, some of the sphalerite exhibits, blebs in exsolved pyrrhotite. The analytical data are shown in Table 3.

Cadmium and manganese were estimated to be below 1% combined CdS and MnS by X-ray fluorescence analysis. Because of the mixed, fine-grained nature of the sulfides, only a single sphalerite was analyzed for copper. This is B 129 at 61 feet, and gives 0.08% Cu. As the data show, very few of the sphalerites contain exsolved chalcopyrite, and hence it is assumed that their copper content is quite low, in keeping with the one chemical analysis.

From considerations of the complex paragenesis and of exsolution relationships, it is obvious that equilibrium between sphalerite and pyrrhotite has been approached but not achieved. As Table 3 indicates, some of the sphalerite is accompanied by pyrite, some by pyrrhotite.

Four pyrrhotites selected for lattice measurement gave an X-ray pattern characterized by the splitting of the d(102) peak indicating that it had inverted to another form. Hence it was not possible to get a confirming temperature from the pyrrhotite-pyrite pair.

TABLE 4

| | | $T_{ZnS}$ | $T_{Po}$ |
|---|---|---|---|
| Heath Steele | Arnold (3) | 550° C | 490° C |
| | Benson (7) | 540 | |
| B. M. S. No. 12 | Benson (7) | 555 | |
| B. M. S. No. 6 | Benson (6) | 555 | |
| | J. K. | 515 | |
| | Arnold (4) | | 440 |
| Clearwater Brook | Arnold (4) | | 470 |

Unknown to the writer, a similar study was undertaken by Benson at McGill University on material from some of the New Brunswick deposits (6, 7). Thus both of us worked on the B. M. and S. No. 6 deposit and unintentionally confirmed one another's results (Table 3). However, the writer draws certain additional conclusions from this data not previously suggested by Benson.

*Discussion of Results.* The iron content of sphalerite tends to be higher when it is associated with pyrrhotite than when it is with pyrite. This relationship obscures the effects of any temperature zonation.

The data also indicate that some sphalerites with exsolved pyrrhotite contain less total iron than other sphalerites which exhibit no exsolution bodies at all. This in itself suggests rather different histories of genesis and cooling for various grains of the same mineral in the deposits.

The minimum temperature as determined from sphalerites, and appropriate for the suite of ore minerals with the exception of the bismuth-bismuthinite pair, is in the hypothermal range of Lindgren's classification. In contrast with this, the regional metamorphic grade and the grade of the wall-rock alteration is no higher than the greenschist facies.

The temperature-depth curves of Fyfe, Turner, and Verhoogen (11) indicate that at a depth of 3 miles, minerals of the greenschist facies would be stable at a maximum temperature of 280° C. If a correction is made for

the sphalerite-pyrrhotite relationship for this depth, the value of 480° C would correspond to 530° C at 3 miles. Hence the maximum temperature indicated by sphalerite exceeds the suggested upper limit of the greenschist facies by 250° C.

Similar high temperatures have been recorded for this and for other New Brunswick deposits, as shown in Table 4. These temperatures are uncorrected for depth of formation. These deposits are rather similar: all are emplaced in rocks of the greenschist facies. The mineralogy might be characterized as representing the hypothermal class. All show an anomalously low grade of wall-rock alteration.

The writer does not know why the high temperatures obtained during the formation of the mineral deposits are not reflected in the mineralogy of the wall rocks. However, he does wish to point out that these anomalous temperature conditions, as well as the complex mineralogy and structure of the deposit, refute the hypothesis that the No. 6 deposit is of sedimentary origin, or is a dynamothermally metamorphosed sedimentary sulphide deposit (20). Surely there would be some evidence remaining in the region had the temperature of the rocks on a regional basis been elevated to 500° C. In view of these temperature relationships, the writer considers this deposit, and others similiar to it, to be of epignetic origin (10, p. 113).

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*July 8, 1961*

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# COPPER, LEAD AND ZINC IN RIVERS DRAINING CHILEAN ANDES

ANN DEGRYS

ABSTRACT

A preliminary geochemical study for copper, lead and zinc was carried out on rivers draining the Chilean Andes between latitudes 34 and 41. Sediments and water samples were collected, the sediments being analyzed for all three metals, and the waters for copper only. The results were plotted as histograms and several areas were found to have above the average metal content. These included: a) The Cachapoal-Tinguiririca-Teno area with copper, which may be due to mineralization present in the area, b) the Chillan district with lead values, and c) the Osorno and Lakes District with high copper and also some lead values. The copper content in this area is thought to be related to the recent basaltic flows present in the area.

Few streams had anomalous zinc values and in general these streams also had high lead or copper values.

PRACTICALLY all the base metal mines of Chile, except for the lead-zinc province of Aysen (lat. 45°45S), occur north of Rancagua (lat. 34°S). There is no marked geological change from Rancagua southwards and therefore no apparent reason why mineralization should suddenly terminate. It is possible that hitherto undetected mineralization may exist south of Rancagua and for this reason a program of geochemical exploration was initiated by the Instituto de Geofísica y Sismología. Drainage sampling was used and the samples analyzed for copper, lead and zinc in order to determine areas of possible interest for more detailed study. The technique of sampling drainage basin has successfully been used in many other parts of the world in mineral exploration programs. It has been used under a variety of conditions, including subtropical Africa for copper (5) temperate regions, such as copper in Eastern Canada (4) and in Arctic regions for copper, for example in the Yukon (1) to mention only a few examples. Basically the principle used is that the erosion products of an area are channeled out through the drainage system and that the rivers carry a load that is representative of the rocks of the area.

## DESCRIPTION OF THE AREA

The Cordillera de los Andes has a core of granodiorite that outcrops extensively south of latitude 38. North of this latitude much of the granite is covered by Cretaceous volcanics, comprising andesites and porphyries and sediments derived therefrom. A line of Recent volcanoes extends along the Andes and there are large areas of basaltic extrusions particularly in the southern part of the region studied.

The Andes is bound on the west by the Central Valley, the change being generally abrupt and commonly marked by a north-south fault line. Both fluvial and/or glacial deposits fill the Central Valley. From latitude 38 southwards between the Cordillera and the Central Valley are a string of lakes that owe their origin to the damming of river valleys by moraines (*2*) and/or by recent movement along faults (St. Amand, personal communication).

The rivers that issue from the Andes are swift flowing with clear waters and sediments ranging from large boulders to fine silt. Volume of the waters carried by the rivers is dependent on the climate. In the northern part of the region the volume increases during the winter rainy season and again during mid-summer due to melting snows. During the summer months the volume is greatly diminished due to evaporation and runoff. Towards the south the seasonal differences are less marked and an increase of the volume of the water due to the rains in winter is compensated in the summer by snow-melt water. Rivers that are fed by the lakes also maintain a constant flow of water throughout the year.

For geochemical prospecting purposes it is advisable to collect samples when the volume of water is at a minimum and offset the dilution effects from the increase due to rain and snow-melt waters. As the volume of the water decreases in the summer, metal content in solution should become more concentrated and lead to the adsorption of metal on to the sediment. Also there are increased proportions of ground water contributions. Once the metal is removed from solution it may not readily be redissolved. It is advisable, particularly in the northern part of the region, to sample during the summer months, but in the south it would make little difference when the samples are collected.

The climate becomes wetter toward the south, and there is an increase of vegetation. The Andean Cordillera round the lake district is densely covered by forest. In recent years there has been a considerable amount of deforestation in progress with the result that surface layers of the soil are stripped off.

## TECHNIQUES

Both water and sediment samples were collected. The water samples were evaporated and then tested for copper content. The sediment samples, after being dried, were sieved through $-80$ and $-200$ mesh screens before analyses for copper, lead and zinc. A nitric acid digestion was used for attacking the samples. The methods employed are based on those in use at the Imperial College, University of London.

In general, the metal content of the $-200$ mesh fraction is higher and these results will mainly be discussed in the ensuing paragraphs.

## RESULTS

Histograms were constructed for the results obtained (Fig. 1) and average metal content for the sediments is 40 ppm copper, 25 ppm zinc, and 25 ppm lead, and 12 ppb copper for the waters. Maps (Figs. 2–6) show the distri-

bution of the metal values. Several areas have higher than average content of metal and are briefly described below.

*Cachapoal Tinguiririca-Teno Area.*—The Cachapoal has the highest copper values both in the sediment and in the waters; the Tinguiririca less and the Teno least. Southwards from the River Teno the values are low, less than 20 ppm in the sediment and 6 ppb in the waters. Some zinc but no lead is associated with the copper.

The well known mine El Teniente is situated on a tributary of the Cachapoal River in the northern part of the Cachapoal drainage basin. Waters drainging this mine are heavily contaminated with copper, lead, and zinc, but on the main branch of the Cachapoal River above the confluence with this tributary, copper values are still anomalous. It is possible that a wide dispersion aureole exists with El Teniente as the focus, and the copper values decrease southwards. This would explain the observed dispersion pattern, i.e., with the highest copper values in the Cachapoal River and a general decrease southwards toward the Teno River. No doubt the aureole extends northwards as well, but the streams to the north have not yet been investigated.

After the general preliminary survey, follow-up work was carried out in the Tinguiririca valley where copper deposits are unknown. The results revealed several "anomalous peaks" and indications of copper mineralization.

Only the streams draining the Andes are anomalous, for those that rise in the Central Valley have low copper values only in the sediment. One stream, the Antivero, has a copper content of 26 ppb (average copper content

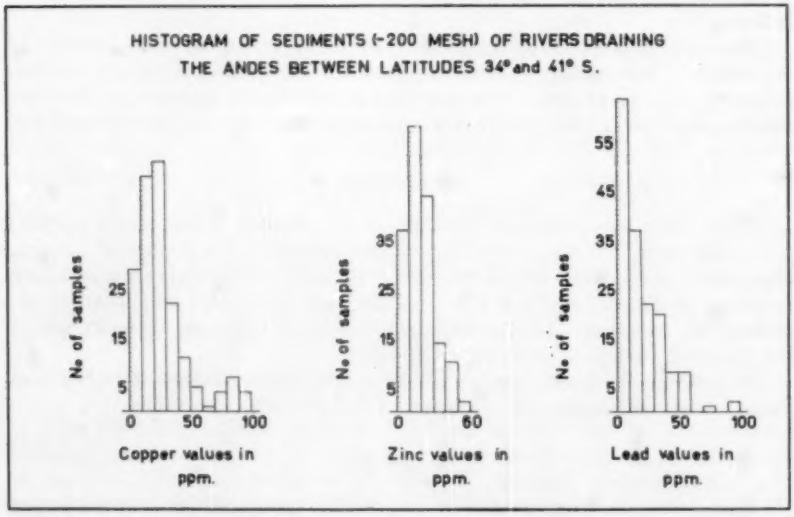


FIG. 1. Histograms of the sediments (−200 mesh) of the rivers draining the Cordillera de los Andes between latitudes 34° and 41°.

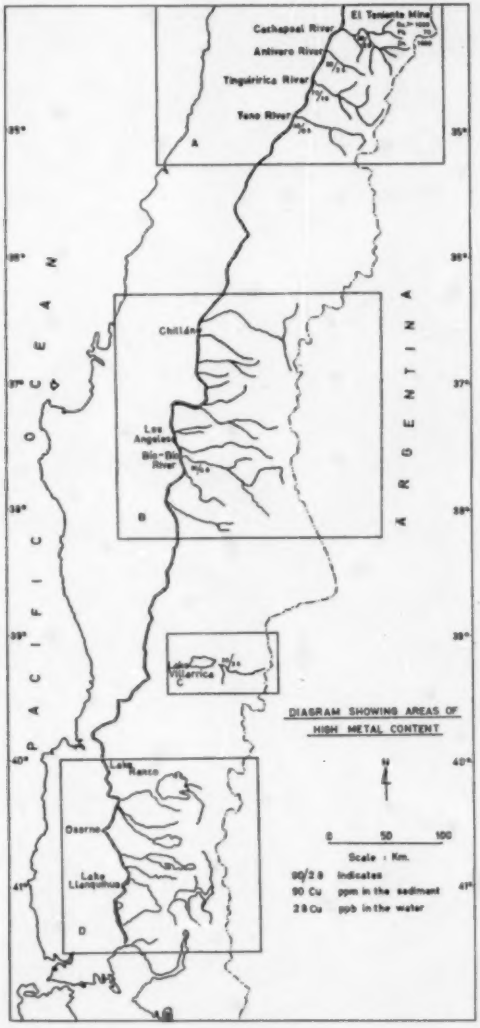


FIG. 2.

for all streams, 12 ppb) in the waters. Microscopic examination of the sediment revealed the presence of chalcopyrite. It is not clear why copper should be present in solution and not in the finer portion of the sediment. It may be that the copper sulfide minerals are eroded mechanically, resulting in fragments present in the coarse sediment and chemically with copper being taken

in solution, but the solutions do not react with the finer portion of the sediment. The clays in the sediment may be saturated, and therefore there is no reaction with the copper-bearing solution. It does not seem likely that the Antivero derives its copper from the Central Valley alluvial fillings, it is more probable that there is some mineralization along the contact of the Central Valley and the Andes where the stream rises.

*Chillán Area.*—Several streams in the Chillán area show indications of lead. No mineralization is known in the area, and in general lead deposits

FIG. 3.

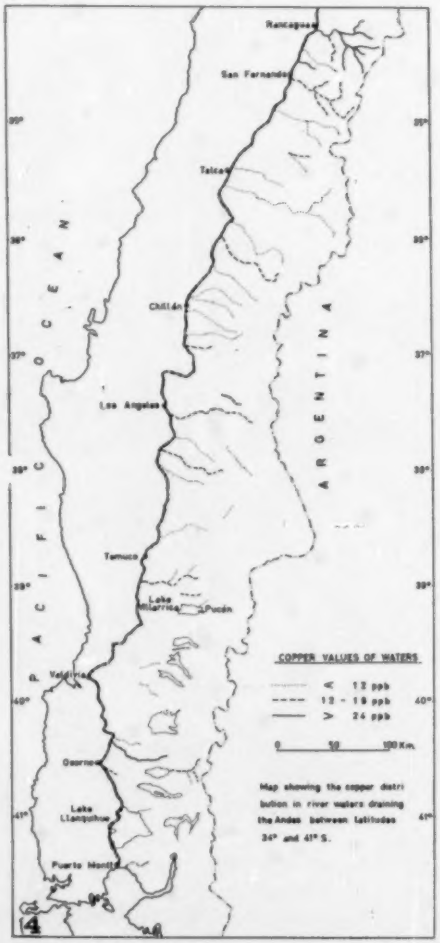


FIG. 4.

FIG. 5.

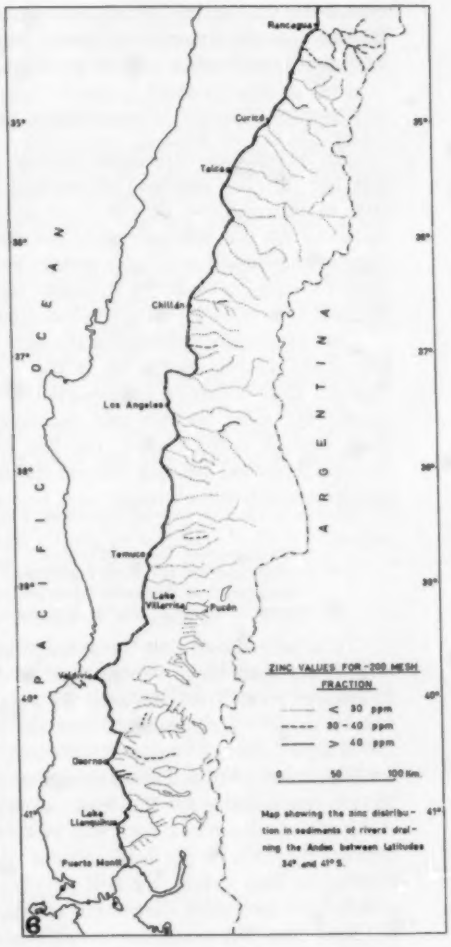


FIG. 6.

are scarce in North and Central Chile. Reasons for the lead values are discussed in the next session.

*Villarrica Area.*—Rivers draining the Cordillera and entering Lake Villarrica have somewhat above the average copper, lead and zinc content. Again, no mineralization is known in the area, but further studies may be rewarding.

*Osorno and the Lakes Region.*—Several streams in this area had above average copper content (50 ppm). Some abnormal zinc values are also en-

countered, but nearly in all cases associated with either the lead or copper. In this area the increases of metal content are not thought to be due to mineralization, and other causes are suggested below.

## DISCUSSION OF THE RESULTS

An increase of base metal content above the average in the rivers may be due to: (1) the presence of mineralization within the drainage area, either sub- or outcropping, that is being oxidized and the oxidation products being removed via the drainage; (2) the rocks in the drainage area have a higher than average content of base metals, and this increase is reflected in the drainage; (3) the rocks in the drainage have only normal base metal content and the sediment is such that it can can accumulate metal from background sources.

As has already been stated the only positive case of anomalous metal values being due to mineralization is the Cachapoal-Teno area. Although in the Osorno Lake District unsuspected mineralization may be present it would appear far more feasible that increased metal content is due to the higher-than-average metal values for the rocks in this area. Numerous recent basaltic flows are present around the lakes, and as shown by the table below, basalts contain more copper and zinc than the average for the earth's crust.

| | Cu, ppm | Pb, ppm | Zn, ppm |
|---|---|---|---|
| Average content for earth's crust | 70 | 25 | 80 |
| Average content for earth's granites | 100 | 5–50 | 60 |
| Average content for earth's basalts | 200 | 7 | 130 |
| Average content for earth's sediments | 40 | 25 | 25 |

This table shows that the average values for copper and zinc in the basalts are higher than those encountered in the Chilean sediments. It is intended to analyze rocks from the area for copper and zinc so that a true comparison can be made. It is possible that the nitric acid digestion used in preparing the samples for analyses is insufficient to release all the metal.

The basalts in the Lake District may explain the copper and zinc increases but do not account for the lead values, either here or in the Chillan district. Analyses of the rocks for base metals would obviously indicate if there is general increase in the lead content of the rocks in the area. Even so, the anomalous lead values are still puzzling since in contrast to copper and zinc, which have extensive dispersion trains, those of lead are generally short and the lead does not move far from the source.

Metal in the drainage may be transported either as part of the sediment or carried in solution by the waters. In general, the Andean rivers have rapid flow and steep gradients and therefore the mechanical transport of the sediment will be an important factor in metal dispersion. However, those elements whose compounds have high solubilities, e.g., copper and zinc, will be transported both by chemical and mechanical means. Metal carried in solution may be removed to the sediment, either by precipitation or base exchange reactions, and become part of the normal sediment load. Elements with low solubility, e.g., lead, will be removed dominantly by mechanical means, but, this is not thought to be the case in the Chilean rivers (see later).

Also affecting the mechanical or chemical migration of the metals is the climate. In the more arid northern part of the region, mechanical agents will play a far greater role in metal dispersion than in the wetter south. In the forest covered lake region where the rainfall is heavy, leaching will be intense. As the total salt contents of the waters is low it may be thought that much of the metal has already been leached out. However, some of the rivers in the southern region do have copper in solution, and a corresponding high amount of copper in the sediment. This suggests that the leaching of the Recent basaltic flows may be responsible for the present metal content of the river waters and sediments.

In the chemical dispersion, metal may be dispersed over a great length, or may become concentrated by some other constituent of the sediment. Iron and organic matter are well known for their capacity to absorb metals. The subsequent follow up work in the Tinguiririca Valley showed that for copper the dispersion train can extend over considerable distance. Dispersion of zinc is probably similar to that of copper. Lead migration in the Chilean rivers is probably as organic complexes. It is obvious that more samples and more analyses are required to give a more accurate picture of the metal dispersion in the Chilean rivers.

The length of the dispersion train is dependent on the size and volume of the stream. If the same amount of metal, either in solution or as part of the sediment, is introduced into a large and a small stream, the values will obviously be lower in the larger stream because of the effect of dilution. It is not always possible to have a complete record of the volume and the velocity for each stream. Therefore, it is likely that for some of the larger rivers the results do not give a true picture of the metal potentiality of the area. This has actually been encountered in the case of the Bio-Bio (these tributaries were not sampled). Values of only 30 ppm copper were obtained for the sediments of the Bio-Bio. To avoid this problem as many of the tributaries as possible were also sampled, but a complete coverage was not achieved. However, the results are felt to give a general picture of metal dispersion in rivers draining the Andes of Central Chile.

## SUMMARY

The results of the reconaissance survey for base metals in rivers draining the Andes of Central Chile may be summarized as follows:

1). A copper anomaly extends south of the Cachapoal River, to the Teno River, probably in the nature of a dispersion aureole with El Teniente as center.

2). High copper values in the sediments were also encountered in the Osorno Lake District. The metal is probably derived from the basaltic flows present in the area. Some high lead values were also encountered in this area.

3). In the Chillán district lead values were higher, the reason for this being unknown.

4). Zinc does not form many anomalies, and where present to any degree, is associated either with lead or copper.

ACKNOWLEDGMENTS

Several members of the Instituto de Geofísica y Sismología assisted the author both in the field and in the laboratory, and their help is gratefully appreciated. Particular mention must be made of Sta. M. E. Arancibia for her laboratory work and excellent draughting of the diagrams. Also, we owe a great deal to the encouragement of our Director, Dr. C. Lomnitz.

The author is also indebted to Prof. J. Muñoz Cristi, Dr. P. St. Amand, Dr. J. S. Tooms, Sres. H. Flores, H. Cusacanqui, and Dr. S. Davis, who kindly offered to review the manuscript and gave helpful criticism and advice.

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*May 18, 1961*

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# SCIENTIFIC COMMUNICATIONS

## *BRIEF OUTLINE OF A POSSIBLE ORIGIN OF COPPER PORPHYRY BRECCIAS*

PAUL KENTS

The two most essential requirements for the occurrence of ore deposition are: a source (mineralizing fluids) and space for deposition (fissures, fractures, etc.). As one of these essentials is a matter of chemistry and the other involves mechanics, it is difficult to explain why these two different domains happen to coincide at the right times and places. This is particularly true in copper porphyries where mineralizing fluids and a specific large-scale shattering of the host rock seem to have come together.

In mineralized districts, whether they are small or large, there generally appears to have been just about the right amount of space available for ore deposition; while in contrast, in surrounding areas no fissures of any significance can be encountered comparable to those that have been mineralized. Thus one may suspect some relation between the extent of fracturing and the volumes of mineralizing fluids involved; or, in other words, it seems as though the mineralizing fluids have themselves provided to a certain extent the space for ore deposition.

Ore deposition can be looked on as a specific stage in a major orogenic geosynclinal development. As elaborated by Vening Meinesz,[1] when two systems of subcrustal convection currents moving in opposite directions meet, a geosyncline is pulled downward between them by the viscous drag of these convection currents. After they are unable to follow the downward movement of the crust any longer, the uppermost layers of the crust are subjected to strong lateral compressive forces, and become folded and overthrusted. After the convection currents cease, the folded geosyncline readjusts itself to the isostatic equilibrium, meaning in effect that the magma will be pressing on the folded mountains from below and lift them up. At the time of orogeny great batholithic masses will invade the folded crustal layers. In the interval between the cessation of convection currents and the beginning of the adjustment to isostatic equilibrium, the cooling magma is relatively undisturbed, allowing it to differentiate. These different magmatic fractions will be injected into the uppermost crustal layers when the equilibrial flow has gathered its full force, and the geosyncline starts to rise. It is during this stage when magmatic pressures from below are at their greatest, when ore

[1] F. A. Vening Meinesz, Plastic buckling of the Earth's crust; The origin of geosynclines: Symposium, Crust of the Earth, Geol. Soc. Am. Special Paper 62, 1955.

deposition occurs; generally associated with intrusives of the most acidic fractions of the cooling magma. As to the source of the copper, or other metals in orogenic geosynclines, most likely they were brought up from the inner depths of the earth by convection currents.

Much has been said and written on the possible processes involved in ore deposition, yet it seems to this writer that the catastrophic explosive volcanic eruptions (Vesuvius, Krakatau, Mt. Pelee, Sakurajima, etc.) have not been examined sufficiently closely to lead to any possible inferences as to the large scale shattering associated with copper porphyry breccias. T. A. Jaggar [2] concludes that these explosions result from ground water turning into steam; however, he does not produce any evidence against a possible magmatic origin of the water involved.

The various manifestations of volcanic activity: effusion of lavas, eruptions of ash and other ejecta, steam and gas emanations, lava flow fissure openings, subterranean movements, upswelling and sinking of lava, etc. are all periodic, each having its own cycle, amongst which 11, 33, 66 and 132-year cycles have been observed. According to observations made in Italy, Hawaii and Japan, the catastrophic steam blast eruptions resulting in the integration of other shorter cycles, occur at intervals of 132 years. Hence, observations at volcanoes indicate that magma is not static, but systematically active: its periodic swellings and sinkings reveal regular pulsations in the magma reservoir. These pulsations, besides influencing observable volcanic activities, are expected to have their effect also on the hydrothermal fluids after these have separated from the cooling and differentiating magma. These hydrothermal fluids, due to their lighter gravity, will rise and accumulate in the highest parts of the intrusive. When magma is swelling, more pressure is exerted against the ceiling of the intrusive, where the most affected constituents are the hydrothermal fluids. These, due to their greater mobility, will be forced in the direction of least resistance, hence, injected upward into the enclosing rock. When magma is sinking, pressures against the ceiling of the intrusive are easing, no further hydrothermal fluids are injected into the enclosing rocks, instead, they accumulate at the ceiling. The succeeding magmatic surge pushes them into the enclosing rock, and drives the earlier fluids already there farther ahead. The general principle involved is crudely comparable to the action of a hydraulic press.

The invaded rocks will yield to magmatic pressures: incipient fractures are wedged open, widened and a host of new ones formed. Laboriously the rock becomes ruptured, pried apart and develops an appearance as if it had been thoroughly shattered. There may be some orientation when some sets of fractures happen to be more consistent and stronger than others. These structures will be referred to below as rupture breccias.

The type of fracture system that will develop depends largely on local conditions such as existing fractures and inherent stress patterns, or structural lineaments, but above all it will depend on the character of the rock affected. A minor feebly mineralized satellite porphyry breccia, about half way between

[2] T. A. Jaggar, Steam blast volcanic eruptions: Hawaiian Volcano Research Assoc., Honolulu, 1949.

the Cuajone and Quellaveco porphyry copper ore bodies in Peru, straddles the contact between monzonite and quartz porphyry. Fractures in the monzonite are completely random, while those in quartz porphyry are far fewer, farther apart, prominent, straight, with lengths measured in scores of feet, and all are parallel and oriented along a single attitude.

The rupture breccia, because of its origin by fracturing, is composed of angular fragments which match with their neighbors. Individual fragments show no movement, dislocation or rotation. Toward the peripheries the intensity of fracturing decreases and the rupture breccia gradually merges into undisturbed rock. This phenomenon is common at numerous copper porphyry breccias.

When during the course of the development of rupture breccia some magma is withdrawn from below, the breccia subsides accordingly and its fragments will become deranged; when the magma rises again the breccia is pushed back, at which process the individual fragments by crushing and grinding lose their angularity, become rounded and assume the character of a conglomerate. If additional amounts of magma were pushed up, igneous intrusives were produced, which in forcing their way ahead took advantage of the trends of structural weaknesses developed.

The conventional hydrothermal theory of ore deposition presumes that upward circulating mineralizing fluids deposited ore according to the decreasing pressure and temperature ranges through which they were travelling. The spent fluids eventually found their way to the surface. It is not known what were the ultimate volumes of hydrothermal fluids liberated from the cooling and differentiating magma, but if their volume amounted to the possible maximum of 9 percent, then there could not have been excessively large volumes of them at any time, and they could have been readily accommodated in the rupture breccias. Accordingly, as the periodic magmatic surges forced in more hydrothermal fluids, there was no outflow: the fluids stayed in rupture breccia, soaking it, depositing sulfides by replacement, and altering the host hydrothermally.

When compared to the ore deposition in fissures by rising circulating mineralizing fluids then, the conditions in rupture breccias are different: above all, the system or environment at mineralization was a closed one, changes in temperature and pressure were relatively insignificant, and ore deposition was effected by replacement. It might be possible to draw a conclusion that confined environment is conducive to replacement, while the circulating flow in open system leads to filling of cavities.

Driven from behind by the incoming hydrothermal fluids, the fluids in ruptures breccia move slowly on, upward and towards the peripheries, and gradually spend themselves. This can explain the ore zoning and the zoning of hydrothermal alteration in porphyry breccias; likewise, it can explain why the orebody at Climax, Colo. has the shape of an inverted cup.

When the composition of succeeding batches of hydrothermal fluids changes, mineral paragenesis will result, as well as different generations of same mineral appearing at different times. If the succeeding batches happen to take a different route, the stress center of rupture breccia would migrate,

also the distribution centers for different sulfides become shifted to different locations. This would also explain why the center of hydrothermal alteration does not necessarily have to coincide with the center of sulfide mineralization.

The most welcome aspect of the confined environment at the site of ore deposition is that, under normal conditions, nearly all of the precipitable sulfides can be expected to remain in the rock, thus explaining the enormous amounts of copper contained in porphyry breccia type deposits; while on the other hand, during the deposition by rising circulating fluids in open systems, there are always chances that some (or most) of the sulfides may escape.

In the process of accommodating the hydrothermal fluids in rupture breccias: the fractures are wedged open, the overall volume expands, and cracks may open in the overlying rocks which would facilitate the upward movement of the confined fluids. If then a magmatic surge happens to be of sufficient strength to overcome the confining pressures, the fluids will break through to the surface. Any sudden release of pressure turns superheated water into exploding steam which will blast a vent through to the surface, manifested there by a steam blast eruption, followed by a water flood of relatively short duration. Afterwards the rubble plugs the vent, leaving a breccia-pipe, like those at Braden and Toquepala in the Andes. Such vents may not extend down deep enough to tap magmatic source, hence explaining the usual absence of intrusive rocks in such breccia-pipes. The usual shape of the vent is circular or oval, but it can also be a fissure, as for example the winding "shear zone" crossing the Climax orebody, which although it has a very impressive width, lacks any apparent displacement, hence may have been the result of a fissure-type eruption, plugged back by ore debris and argillized afterwards.

Breccia pipes besides accompanying a number of copper porphyries, are also found in the vicinities of some of the fissure-type vein deposits, indicating that perhaps at least in some cases, a confined environment might have existed during the earliest stages of the development of fissure-type deposits. The opening of fissures and wide hydrothermal halos around them may have originated at an earlier confined stage; and after the confinement had ceased, the fractures were filled by rising circulating hydrothermal fluids. Also, in rupture breccias, after the confinement had eased, mineralization by replacement was succeeded by cavity filling whereby the fractures became healed by sulfides and quartz.

Large scale hydrothermal alteration extending for considerable distances away from the vein systems may be perhaps the most characteristic indication of confined environment. No other plausible explanation has yet been offered to explain the mechanics involved in such wide penetration by hydrothermal alteration.

The aim of this communication has been to introduce a thesis as to the probability of confined environment existing at the replacement type of hydrothermal ore deposition, embracing the many varied but seemingly related phenomena at ore deposition, and fitting them together into a comprehensive genetic sequence.

The concept of enclosing rocks yielding to hydraulic pressures, produced by magma and transmitted to the site of fracturing by hydrothermal fluids, though perhaps novel, is amply supported by field evidence, and in this writer's experience, rupture breccias, dendritic and horsetailing fractures, and steam blast vents are rather common in Peru; but, as in most instances these do not contain any desirable mineralization, little or no attention has hitherto been paid to them.

CASILLA 197-D,
SANTIAGO, CHILE,
*Sept. 28, 1961*

## *URANIUM-BEARING BODIES IN THE ORE DEPOSITS OF TRAVERSELLA (ITALY)*

STEFANO C. ZUCCHETTI

The Mining Department of the Società Fiat, upon the discovery by their mining experts of radioactive phenomena in some types of mill concentrates in the Traversella mine (exploited for magnetite, scheelite and various sulfides), has entrusted the "Sezione Miniere" of the Turin Polytechnic School with a geologic-mining study of the deposit, in order to identify any radioactive minerals present therein, to define their paragenesis and distribution in the mine and to supply useful guide elements for establishing their consistency, as well as the methods for their possible concentration.

The ore deposits of Traversella lies in northwestern Piedmont and more precisely in Turin province. It is a characteristic contact-metasomatic deposit at the contact between granodiorite and pre-existing enclosing rocks (crystalline schists of the "Sesia-Lanzo" massif).

In a particular hornfels facies of the thick and variable aureole, which forms the productive area, in part exploitable, of the mine, I found the radioactive mineralization. The radioactive mineral is uraninite, thoriferous type. It occurs in isometric euhedral crystals, in polyhedric forms, both alone and in combination. The microscopic examination of many scores of crystals of particularly well defined and developed habit, shows the following forms, both simple, isolated or associated to one another: cube, octahedron, dodecahedron, tetrahexahedron. The maximum dimensions of the single polyhedra are of the order of 1 mm or little more. The average specific gravity of the uraninites of Traversella is near 9.

In transmitted light, this uraninite appears opaque, even in very thin sections. In polished sections it has a gray color, tending to brownish. The reflecting power, measured with the Berek's microphotometer, is between 14 and 16 percent.

We have also made a complete quantitative analysis of this uraninite using thousands of grains from some mill concentrate.[1] The results obtained are:

[1] The analysis, followed by myself in all its stages, has been performed by E. Matteucci, a chemist of the "Sezione Miniere."

| | | | |
|---|---|---|---|
| U | = 79.03% | W | = 3.07 |
| Th | = 0.37 | Fe | = 0.49 |
| Pb | = 0.47 | Mn | = 0.08 |
| Bi | = 0.34 | Ca | = 0.47 |
| Cu | = 0.12 | S | = 0.42 |
| Au | = traces | Si | = 0.51 |
| Mo | = 0.06 | $O_2$ | = 15.87 |
| | | Total | = 101.30 |

As appear from the analysis data, it has not yet been possible to separate completely the uraninite from grains of the other minerals (such as wolframite, molybdenite, chalcopyrite). Uranium has been—as a first approximation—considered in the form of $U_3O_8$. Further research work is being made in order to determine the degree of oxidation of $U^4$ to $U^6$.

The presence of uraninite in the mill concentrates has been confirmed by X-ray analysis. The roentgen methods have further confirmed the presence of native bismuth and gold, identified under the microscope.

The micrographic and petrographic studies carried out on several tens of polished and thin sections, obtained from many samples collected underground, yields the following conclusion:

1. In the deposit the uraninite is associated with other metalliferous minerals, mainly sulfides, i.e., molybdenite, pyrrhotite and chalcopyrite, and only subordinately with wolframite, scheelite, magnetite, pyrite and arsenopyrite;

2. The uraninite occurs always and only in those facies of the metamorphic aureole rich in dark silicates, such as brown mica, chlorite, amphibole, pyroxene, serpentine, olivine, or talc, with which are associated also carbonate and quartzose matrices, in subordinate amounts;

3. In the uranium-bearing hornfels, the light silicates—such as garnet, plagioclase, zoisite, gehlenite, wollastonite, etc.—always occur as accessory minerals, or are lacking. Where they prevail, for instance in all the mineralized garnetites, the uraninite is absent, and there is no trace of radioactivity;

4. Among the metalliferous minerals present in the uranium-bearing hornfelses, uraninite is always one of the products of first crystallization. In fact, as already pointed out, it occurs as euhedral crystals, commonly included or replaced by other minerals, both metallic and nonmetallic.

Detailed prospecting of the areas of highest radioactivity has permitted the distinction, inside the metamorphic aureole, of those facies with ferromagnesian silicates, mineralized by uranium, as described above. They form distinct autonomous bodies, which may be distant from one another. Their distribution in the aureole is quite irregular and their form and dimensions are also irregular. In fact, we find bodies of rounded or lenticular shape, and others of columnar shape, with variable thickness; some shoots have ragged ends or branch and twist.

In the mineralized uranium facies, the uraninite is distributed in the same manner as the other minerals. Uraninite, like molybdenite, pyrrhotite, chalcopyrite and the other metallic minerals present in these hornfelses, represents one of the products deposited by the magmatic fluids, which also brought in the nonmetallic minerals of late formation. Finally, this uraninite is con-

temporaneous with the metamorphic and metallizing activity that caused the formation of the hornfelses, of which it represents a constituent, in small proportions.

The uranium deposit of Traversella must be classified as contact metasomatic in a perimagmatic area.

E. W. Heinrich (1958, Mineralogy and geology of radioactive raw materials) has classified these contact-metasomatic deposits, based essentially upon their different mineralogic associations. According to this classification, the radioactive mineralization of Traversella can possibly be placed in the subtype no. 1 (deposits of thorianite-uraninite in marble, etc.) or also, if one considers the deposit as a whole, in the subtype no. 4 (deposits of magnetite, with uraninite, etc.).

POLITECNICO—SEZIONE MINIERE,
TORINO, ITALIA,
*Sept. 3, 1961*

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# DISCUSSIONS

## *GOLD-BEARING LODES, KOLAR, INDIA*

*Sir:* M/S Narayanaswamy, Ziauddin and Ramachandra are to be congratulated for their excellent paper "Structural Control and Localization of Gold-Bearing Lodes, Kolar Gold Field, India" (Vol. 55, No. 7) on one of the important gold mines of the world, based on data meticulously collected in the field and intelligently interpreted. The field work was taken up, as professed, with a view to locate "new lodes and for tracing extension of the known lodes" because of the gradual "depletion of the high-grade ore reserves." Towards this end, from the data presented in the paper, the following observations can be made:

1) Contrary to views previously held, the gold lodes do not appear to be genetically connected with the *Tirlapalle* laccolith of Champion gneiss, the intrusion of which structurally affected the lodes, which now echo the northern contact boundary of the laccolith.

2) The lodes were, therefore, formed prior to the intrusion of the Champion gneiss laccolith.

3) The gold field owes its preservation from erosion due to the number of oblique faults that traverse it. The Balaghat North fault alone appears to have a throw of 9,000 feet to the south-west (Fig. 2). Similarly the Mysore North fault also throws the lodes to the south-west. Due to long erosion the schist belt occurs in attenuated outcrops to the north and south of the field; and had the faults not taken place, the productive portion of the field would long have been eroded away.

4) The faults are, however, later than the intrusion of the Champion gneiss, as the Mysore North fault has affected the Champion gneiss.

5) In the area north of the Balaghat North fault, where the root of the syncline has been exposed, the lodes are of poor grade and the workings there have extended only to a depth of 600 feet.

6) It, therefore, follows that the grade of ore in the working lodes will fall as the roots of the syncline are approached in the productive portion of the field. In other words, the lodes will become non-paying like the area to the north of the Balaghat North fault as the workings reach the full extent of the combined throw of the fault. The exact determination of the throws of the faults is, therefore, of utmost economic importance.

S. K. Borooah
U. C. Hazarika

Raipur, India,
*Oct. 13, 1961*

## PALEOZOIC BANDED IRON-FORMATIONS

*Sir:* The value of J. E. O'Rourke's paper will not be lessened by deleting his example (p. 349) headed "The Setolazar District of Morocco," derived from the garbled account of Heim.[1] A better description was provided by de Lizaur.[2]

Briefly, the deposit of Setolazar is a mid-Tertiary metasomatic replacement of lower-Cretaceous limestones and intercalated limestone-shales by magnetite and pyrite. The top 20 to 40 m of each outcrop weathered to hematite. A fuller account of all the iron-ore mines in NE Morocco is to be published shortly.

H. N. RHODEN

MELILLA, SPAIN,
*Aug. 14, 1961*

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2. de Lizaur y Roldan, J., 1952, Gites de Fer des Kelaia: 19th Int. Geol. Cong., Algiers, Livret-guides au Maroc no. 7, p. 24–27.

## CHEMICAL CHARACTERISTICS OF THE GROUND WATER RESOURCES OF JAMAICA, W. I.

### A REPLY

*Sir:* Versey's criticism of our description of one of the aquifers is based on the tacit assumption that we stated or implied that this aquifer is composed almost entirely of clays. This is certainly not the case, and in fact our description of it does not differ from his which is given below. He describes the "heterogeneous alluvium deposits of the plains of Vere as being very variable both vertically and horizontally, the rapidity of the changes making it almost impossible to correlate the strata in the bore holes. Noteworthy are the Savannah or Heyes gravels which occur in a band up to a mile wide, stretching north to south from Hanbury to Raymonds; the 'Marine clay,' which is extensive and of considerable thickness in the 'buried valley,' and the *terra rossa* which occurs in addition to its outcrop over the limestone surfaces, as a narrow fringe to the plain" (5, p. 10). They further state that "clays are very extensive in the areas where the limestone is at depth and in many places they prevent or retard the upward or downward movement of water" (5, p. 16). Our work showed that kaolinite and montmorillonite are the principal clay minerals present. We find it difficult to understand how he finds our description "peculiar." His is lengthier, but certainly no different from our own.

We are continuing a detailed programme of chemical analysis and find that the water from several of the wells, particularly the alluvium wells sometimes contain high concentrations of nitrates and other nitrogen bearing materials.

This we believe to be due to the seepage of rain and irrigation waters downwards through heavily fertilized fields, particularly by nitrogen fertilizers. Our experience is that when the nitrogen content of a water is high the balance of the ionic equivalents is not satisfactory but the reason for this is not yet fully understood; Hem (2) remarks that "the concentration of nitrates in water does not seem to have any consistent relation to the concentrations of the cations."

The scatter in the plot of total solids is partly due to variations in nitrogen-bearing materials that we did not include in our analysis because our facilities do not present nitrogen analyses. Another factor, but a minor one, is that we plotted total solids in ppm versus conductivity in micromhos/cm². Rossum (4, p. 631) shows that the conductivity value for ppm varies from 0.71 for bicarbonate to 3.82 for magnesium, and so we would expect a tendency for scatter if the composition is varying rather than being simply diluted as is usualy the case where one well or stream is studied over a period.

We do not propose to discuss here at length the relative merits of the use of histograms and unique frequency curves. In fact the latter method has long been accepted as a very sensitive test of the homogeneity of populations as particle size in sedimentary petrology (1, 3). Versey's use of histograms seems to us to be an unfortunate choice for these types of comparisons because it is dependent upon the selection of class intervals. On the other hand the unique frequency curve is independent of the particular class interval selected (3).

Versey has provided the best proof of the discriminating sensitivity of the method. In his letter he admits of two types of ground water present in the area. He has also further stated that the "buried valley" contains a thick sequence of sediments including clays and silts that were deposited in salt water. As many of these marine clays have never been above sea level, and because their permeability is low, it is probable that the water contained in them is more concentrated than water that occurs in coarser grained deposits (5, p. 26). Therefore, it is only to be expected that this connate water influences the chemical quality of the water derived from wells drilled into the aquifer. The influence of sea water is also an important factor. An examination of our paper (p. 534) shows that we have been able to detect the presence of what we call our 'B' zone, that is the water derived from the "buried valley," and explains the discontinuity that we have determined, rather than being guessed at as being a gradual change. The fact that the conductivity values, and sodium and chloride content of the samples all give the same results is further evidence that the conclusion is not based on chance, but is real. These values are direct instrumental determinations and are also the ones that would be expected to be most sensitive to differences caused by intruding sea and connate waters intermixing with shallow ground water.

The question of the piezometric surface is answered in Plate I of the paper by Versey and Prescott (5). We however agree that the figure we gave for the maximum observed depth of the alluvium should have been 750 not 400 feet.

We should also remark that this paper is only a preliminary report be-

cause we consider the application of the statistical technique novel and we hope to provide much more detail on the subject in future papers. The problems of hydrochemistry under tropical conditions is interesting because of the differences that can occur due to the effect of the higher all year temperatures on the possible chemical reactions.

V. C. Hill and A. C. Ellington,
Scientific Research Council,
Kingston,
*Oct. 26, 1961*

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# REVIEWS

**Geology of Ore Deposits** (Geologiya Rudnykh Mestorozhdenii). Academy of Sciences of the USSR. January–February 1960; Vol. 2, No. 1 (in Russian).[1]

N. M. STRAKHOV, *Climate and the accumulation of phosphates* (Pp. 3–15, table).

The distribution of Cambrian, Permian, and Tertiary phosphate deposits throughout the world indicates that 80 percent of all phosphorite reserves accumulated in marine basins of arid zones, including such deposits as the Permian Phosphoria formation in the United States and the Lower Paleozoic phosphate deposits of Kara Tau in Turkestan.

The accumulation of phosphates in arid zones results in formation of high-grade oolitic and fine-grained bedded deposits in association with dolomites and gypsum. Under these conditions the high concentration of phosphorus was the result of sedimentation in platform regions, geosynclines, and foredeeps. On the other hand, the low grade concretionary phosphates are typical of the humid zones and are located almost exclusively on the platforms.

A. V. Kazakov suggested that the sedimentation of phosphates resulted from the ascending cold deep sea currents in the shelf zone under conditions of lower partial pressure of carbon dioxide and high pH. Strakhov modifies this point of view by indicating that formation of phosphorites in intracontinental marine basins may be ascribed only to the concentration of $P_2O_5$ in deep horizons of the seas.

The role of the underwater hydrothermal source of phosphate is similar to that of phosphate derived from the continents during weathering. Both sources serve to replenish the total amount of phosphorus in sea water.

M. I. ITSIKSON, V. S. KORMILITSYN, L. I. KRASNYI, AND V. T. MATVEYENKO. *Main features of metallogeny of the northwestern part of the Pacific ore-bearing belt* (Pp. 16–44, 2 maps).

The northwestern part of the Pacific ore-bearing belt includes the eastern Transbaikal region, the Far East, and the Northeastern USSR. The belt consists of the internal zone, which borders the ocean and is characterized by a high concentration of copper, and the external zone, adjacent to the internal zone and containing abundant tin and tungsten deposits. The difference between the two zones is explained by the difference between the continental and oceanic structures.

In the Eastern Transbaikal region the Paleozoic metallogeny is represented by contact-metasomatic iron deposits. The Late Jurassic metallogeny is characterized by the deposits of tin, tungsten,[2] molybdenum, gold,[3] lead, zinc, and probably fluorine. During the Cretaceous metallogeny, the predominant mineralization was represented by gold and tungsten, localized in areas of Jurassic metallogeny. The role of antimony and mercury increased considerably during the Cretaceous period. Rare metals mineralization is associated with the axial part

[1] Regular preparation of reviews of this journal is supported by a grant from the National Science Foundation.

[2] The tin-tungsten ore-bearing belt is 1,000 km. long.

[3] The association of gold and molybdenum is very typical of Jurassic metallogeny.

of the geosyncline, while gold, molybdenum, lead, zinc, antimony, and mercury represent the mineralization inherent to the marginal part of the geosyncline. The southern part of the Far East is characterized by gold placers associated with Precambrian metallogeny, Sinian-Cambrian iron, manganese, and magnesium sedimentary-metamorphic deposits, and endogenous deposits of fluorspar, lead and zinc. Little tin, tungsten, and molybdenum mineralization developed during early Cretaceous time. There is a transition of tin mineralization into gold-bearing veins, mainly within the regions of subsidence.

The sulfide mineralization of Neogene age is developed in the Ust Amur region and on Sakhalin Island. It is accompanied by silver and gold deposits, as well as by some mercury mineralization. The authors suggest a similarity of this part of the Far East with the gold-bearing provinces of the western United States.

In the northeastern part of the country the Paleozoic metallogeny is represented by occurrences of native copper and copper sulfides in garnet skarns. The mineralization of Late Jurassic age corresponds to the period of transformation of the geosyncline into a folded belt and is mainly represented by gold, tin, and tungsten. The Late Cretaceous mineralization corresponds to the period of consolidation of the orogenic belt and to its transformation into a platform. It is of great practical interest and is represented by cassiterite-chlorite-tourmaline and cassiterite-sulfide deposits. There are also lead, zinc, and mercury deposits.

The metallogenetic provinces are characterized by considerable concentrations of tin, gold,[4] lead, and zinc, fluorspar, molybdenum, and to a lesser extent, by tungsten and antimony. Very typical is the association of gold and tin mineralization. The oceanic provinces contain deposits of gold, silver, mercury, native sulfur, barite, and some lead, zinc, aluminum, platinum, and chromium. The mineralization of the continental provinces belongs to the middle and late stages under conditions of weak magmatism. On the other hand, the mineralization in the oceanic group of provinces is related to ultrabasic, basic, and mesocratic magmatic rocks, and corresponds to the incipient stages of the orogeny.

The pre-Mesozoic mineralization within the northwestern part of the Pacific belt is of secondary importance. Almost the entire mineralization developed in the continental group of provinces during the Late Mesozoic and Early Paleogene, and in the oceanic provinces—mainly during the Neogene, and to some extent during Upper Cretaceous and post-Pliocene time. In each metallogenetic province, there is a tendency for the formation of younger endogenous ore belts in a direction away from the center of the orogenic region and toward the margin of the latter. This reflects the evolution of the Pacific region geosynclines which are accompanied by activation and the formation of deeper and more differentiated magmatic chambers to which mineralization is related.

The distribution of metallogenetic regions depends on extensive and long existing fractures of lineament type. Data available on the regional faults indicate that they continue to depth of 700 km and dip 32 to 60° toward the continent. The continental block is overthrust over the oceanic platform along the marginal depression associated with these faults.

Faults of planetary magnitude and metallogenetic importance exist during long periods of time, correspond to the contacts of major structures, such as orogenic regions and shields, and determine the formation of ore-bearing provinces (gold, tin, etc.). The next group of regional faults of metallogenetic importance forms at the contacts of structures within the same orogenic region and are characterized

[4] The gold-bearing belt of the Amur region is 1,500 km long and probably continues through the Kolyma depression and Chukotka to Alaska and British Columbia, forming the 10,000 km-long North Pacific gold-bearing arc.

by batholithic granite intrusions accompanied by quartz veins or skarns containing tin-tungsten and rare metals, and by gold-bearing quartz veins. Faults transverse to the orogenic region are associated with late stages of the evolution of folded structures. These faults are accompanied by hypabyssal and subvolcanic mesocratic and acid intrusions with tin sulfide and lead-zinc mineralization. Relatively short faults formed during the late stages of consolidation of the folded structure along the grabens or depressions. These faults are accompanied by dikes and by fluorspar, gold, antimony, and in some cases, by epithermal gold and antimony-mercury-tungsten mineralization.

A. I. GINZBURG AND G. G. RODIONOV, *The depths of formation of granite pegmatites* (Pp. 45–54).

The depth of formation of pegmatites is determined as a function of the thickness of overlying sedimentary formations, degree of metamorphism of the country rocks, mineral composition, and the gaseous-liquid inclusions. The study of these inclusions by Kalyuzhny indicates the pressure during the formation of quartz to be 220 to 430 atm.

The information on pegmatites of different facies is summarized in the following table:

| | Depths of formation | | | |
|---|---|---|---|---|
| | Below 8 km | 6–8 km | 4–6 km | Less than 4 km |
| Economic types of pegmatites | Rare earths-bearing (monazite, allanite, and uranium-bearing) | Muscovite and feldspar | Rare metals (tantalite, beryl, spodumene, lepidolite) | Rock crystal, optical grade fluorspar |
| Age | Archean | Precambrian, Caledonian, Hercynian | Different | Mesozoic |
| Parent intrusions | Alaskite microcline granites | Biotite-microcline-plagioclase pegmatites | Biotite-microcline-plagioclase granites | Mica granites |
| Facies of metamorphism | Granulitic | Staurolite-kyanite | Cordierite anthophyllite | Green schists, not metamorphic rocks |
| Structure | Massive | Massive, zoned | Zoned | Zoned |

The role of metasomatic processes increases with depth due to the increase in volatiles and alkalies (Na, Li, Rb, and Cs). Deposits of tin, tungsten, and molybdenum appear in the regions of rare metal-bearing pegmatites but are not related in space with the latter. In higher horizons, however, there is an association of pegmatites with deposits of these metals.

F. I. VOL'FSON AND S. I. LEZIN, *Main features of lead zinc deposits of the Elbruz ore field* (Pp. 55–69).

The deposits are located in the Northern Caucasus in the basin of Kuban River. The country rocks are metamorphic formations, probably of Paleozoic age. Mineralization is of the disseminated type in brecciated Late Variscan dikes of quartz porphyries. Disseminated ore and ore of finely banded texture occur also in meta-

morphic schists which are, in general, less favorable for localization of mineralization. The mineralization developed in five stages in connection with the tectonic movements along the fractures. Each stage was characterized by the following minerals: (1) sphalerite with some pyrite and chalcopyrite; (2) pyrite and arsenopyrite, associated with quartz; (3) sphalerite, galena and calcite; (4) calcite veinlets; and (5) realgar-orpiment veinlets. The ore bodies have the form of ore shoots, single, and compound veins, and ladder veins. Ladder veins were less favorable than ore shoots located at the intersection of two systems of fractures for the formation of economic ore grades.

A. V. Druzhinin, *Some features of the geological structure of the Bukuka deposits* (Pp. 70–82).

The quartz-tungsten deposits of Bukuka are located in the Transbaikal region. A 300 to 600 m thick belt of silicified Jurassic sediments in the roof of the Cimmerian granodiorite massif separates the area into western and eastern parts. Dikes of granodiorite-porphyrites and lamprophyres, accompanied by flat, steep quartz-tungsten veins and stockworks are characteristic of the western area. There are no dikes in the eastern part of the field which is represented evidently by granodiorites which are younger than those of the western part. The mineralization is represented by steep dipping galena-bearing veins filling open fractures. Bismuthinite and molybdenite are characteristic of the flat lying veins. High grade tungsten ore in the latter veins forms courses in bends of the fracture which are oriented with their convex sides upwards. The contact of veins with lamprophyre dikes is an example of favorable lithologic control of tungsten mineralization. The structure evolution of the area involved five periods of faulting and fracturing.

E. M. Nekrasov, *Conditions of localization of enriched parts of ore bodies in the deposit of Uch-Ochak* (Pp. 83–91).

The lead zinc deposit of Uch-Ochak is located on the southern slope of Kuraminsky range in southwestern Karamazar, in Tadzhikistan. The ore-bearing veins occur in Lower Permian arkosic conglomerates and sandstones with diabase and labradorite porphyry dikes. The earliest veins consist of quartz with some pyrite and arsenopyrite. These veins are intersected by galena- and sphalerite-bearing quartz veins. Both types of veins are in turn intersected by barite veins. Most of the ore shoots formed in bends of the ore-bearing fractures striking 65 to 70° NE. The ore shoots form also at the junction of fractures and at the sites where the ore-bearing fracture meets the transverse fault or dikes of basic rocks.

V. I. Rekharsky, *Behavior of pitchblende under the influence of fluorine-bearing hydrothermal solutions* (Pp. 92–97).

During the study of an unidentified uranium deposit, the author examined the intersections of uranium-bearing veinlets with the younger fluorspar veinlets within the mineralized zone in acidic, hydrothermally altered intrusive rocks. Pyrite, molybdenite, galena, sphalerite, and chalcopyrite form mainly the central parts of uraninite-bearing calcite veinlets. Uraninite has a colloform structure with sulfides filling the desiccation fissures. Uraninite, calcite and quartz are corroded at the intersection of the uraninite-bearing and fluorspar veinlets. At distances up to 20 cm from the intersection fluorspar contains small inclusions of radioactive minerals which are probably uranium oxides or fluorine-bearing uranium complexes. These minerals occur together with colloform molybdenite, as well as inclusions of galena, pyrite, and chalcopyrite. The fluorine-bearing hydrothermal solutions leached and redeposited the latter minerals mainly in veinlets joining

the fluorspar veinlets. U, Mo, Fe, Pb, Zn, Cu, Ca, and Si could be transported by the fluorine-bearing solutions in some cases to a considerable distance.

R. P. Rafal'sky and Yu. M. Kandykin, *Some experimental data on the crystallization of uranium dioxide under hydrothermal conditions* (Pp. 98–106).

The hydrothermal ore-bearing veins could form not only during slow crystallization from saturated solutions, but also because of the strong supersaturation which resulted in precipitation of finely dispersed sediments. The crystallization of $UO_2$, as the result of reaction of hydrogen sulfide with uranyl sulfate and uranyl-carbonate solutions at temperatures of 280 to 320° C and high pressure, was studied by means of an electron microscope and X-rays. The carbonate solutions are not favorable for crystallization of uranium dioxide. $UO_2$ formed from these solutions has a very imperfect crystalline lattice. Presence of NaCl and especially of NaF contribute greatly to the process of crystallization. The sulfate solutions are more favorable for crystallization of $UO_2$, which sometimes forms globular particles with more or less perfect crystalline structure. The presence of small amounts of NaCl and NaF has no substantial effect on the process of crystallization from sulfate solutions.

V. N. Moskaleva, *The problem of jadeite* (Pp. 106–113).

The author reviews the existing hypotheses on the origin of jadeite as well as experimental studies on the synthesis of jadeite. The point of view concerning the origin of jadeite as a metasomatic product of desilication of albite is supported by the results of study of jadeites from Lake Balkhash region and Upper Burma. In samples from the first region, relics of albite occur between large crystals of jadeite. In other cases secondary albite corrodes the crystals of jadeite, which is present also as relics within albite. The jadeite rocks are hydrothermal-metasomatic formations which formed as the result of deanorthitization and desilication of granitoid dikes in ultrabasic massifs. The pressure of jadeite synthesis was evidently not higher than that which occurred during the formation of ultrabasic intrusions within the folded regions.

P. K. Vinokurov and E. S. Gutkin, *Problem of the zone of weathering and its relation to bauxites of the Northern Urals basin* (Pp. 114–119).

The zone of weathering in magmatic and pyroclastic rocks of Silurian age is up to several meters thick and shows gradual decomposition of feldspars and the formation of chlorite, hydrous mica, halloysite, and kaolinite. The weathered porphyry contains magnetite and titanomagnetite which also occur in bauxites of the basin. This is considered as an indication that the redeposited bauxites formed from the above mentioned products of weathering. Native copper is present in porphyries. It is thought that it may also be responsible for the presence of malachite in bauxites.

S. I. Naboko, *The first All Union volcanological conference* (Pp. 122–125).

The conference took place in Erevan from September 23 to October 2, 1959. Three hundred specialists from sixty-eight organizations (branches of the Academy of Sciences, Ministry of Geology and Conservation of Mineral Resources, and universities) participated at the conference. There were one hundred and five papers presented, followed by sixty-seven discussions, dedicated to: (1) the problems of active volcanism and principles of volcanology; (2) volcanic provinces and formations, and the related mineral deposits; (3) volcanism and tectonics.

Eugene A. Alexandrov

Department of Geology,
Columbia University,
*Sept. 5, 1961*

## *SHORT REVIEWS OF NEW RUSSIAN BOOKS*

**Osnovy Geologii SSSR.** [Principles of the Geology of the U.S.S.R.] By O. L. EINOR. In two volumes. Vol. I, 337 pp. University Press, Kiev. 1960. Price 9r. 80k.

Of the many textbooks on the geology of the U.S.S.R. published in recent years this work from the University of Kiev, so far as it is yet available, is one of the most lucid and most readable. The first of two volumes contains an introductory section giving a historical résumé of geological investigations in Russia (18 pp.), followed by a statement of the principles and terminology of stratigraphic-tectonic classifications (24 pp.); and the rest of the book reviews the regional geology of European Russia, including the Urals and their arctic extensions, the Caucasus, and the Carpathians, within four main chapters concerned with pre-Cambrian, early Palaeozoic, Hercynian and Alpine zones of folding. Presumably the second volume will concern itself with the geology of the Asiatic territories. As illustrations there are 135 instructive sketch maps and sections, not all of which are well drawn. The work is essentially a teaching text rather than a reference book, and the simplifications introduced to make its message more readily intelligible to students will be equally acceptable to most Western readers.

**Osnovy Teorii Litogeneza.** [Principles of the Theory of Lithogenesis.] By N. M. STRAKHOV. In three volumes. Vol. II, 576 pp., 221 figs. Academy of Sciences, Moscow, 1960. Price 36r. 50k.

The first volume of this work has already been briefly noted, and the prognostications of its scientific importance have been endorsed by the news that it has brought the author a Lenin Prize. The second volume reviews the conditions of deposition, diagenesis, and lithification of sedimentary rocks under all kinds of humid conditions, and the third will deal with sediments formed in an arid environment. The immense amount of new analytical data, all fundamental to a proper understanding of lithogenesis, contains much that is relevant to arguments on the origin of ore deposits. Some geologists in the Rhodesian Copperbelt, for example, who seek analogues of zonal ore-bodies in euxinic sediments, would do well to consult the many maps showing the distribution of trace elements in the bottom deposits of the Black Sea, maps which completely discredit such syngenetic hypotheses. It would be presumptuous to criticize Dr. Strakhov's very fine monograph, which illustrates the strength of Russian geology in this important field; but it is surprising that no reference is made to the important part played by tectonic conditions such as anticlinal uplift in governing the intrastratal solution of minerals and the redistribution of their component elements, so clearly demonstrated by Smithson, for Jurassic sediments in Britain, some twenty years ago. [The zonation of Copperbelt-type ore-bodies can more readily be explained by processes of this kind.] An English translation of the work is in hand.

**Yadernaya Geokhronologiya.** (Nuclear Geochronology.) By I. E. STARIK. 630 pp. Academy of Sciences, Moscow, 1961. Price 3r. 13k.*

A frequent feature of Soviet publishing is the remarkable rapidity with which scientific works are printed. This well-produced volume went to press on 3rd

* All books published after 1st January 1961 are priced in new roubles, ten times the value of the old rouble.

November and printing was completed by 31st December 1960. The author, who is Director of the Khlopin Radium Institute in Leningrad, has set out to review the theoretical issues and practical techniques surrounding the determination of the absolute age of rocks and minerals by radioactivity methods, a field in which he has been active for many years. His work is the fullest text on this topic yet to appear and is especially noteworthy for the attention it gives to problems of the possible natural fractionation of radiogenic isotopes, particularly the isotopes of lead. Using Russian data a new time scale is created for the geological column, this agreeing well with the new scales of Holmes and Kulp apart from allocating only 20 million years to the Silurian (420–400 m.y.) and placing the base of the Cambrian at 570 m.y. Four divisions of the pre-Cambrian are recognized: I (Riphean = Proterozoic I), beginning at 1100–1200 m.y.; II (Proterozoic II), beginning at 1800–1900 m.y. [but the Witwatersrand System is older than this!]; III (Archaean), beginning at 2000–2200 m.y.; and IV (Katarchaean), from 3400–3500 m.y. Although most of the book is based on the world-wide researches of the last five years, progress in this field is so rapid that the work will doubtless be obsolescent by the time a translation can be produced.

**Uranovykh Mineralizatsiya Rudnykh Gor.** [Uranium mineralization of the Erzgebirge.] By Yu. M. Dymkov. Pp. 100. Atomizdat, Moscow, 1960. Price 3r. 40k.

This account of the uranium mineralization of the Erzgebirge (Krušné hory) is based entirely on published works, principally German and Czech papers dated from 1860 to 1959. Notwithstanding the extensive postwar mining activity in Bohemia and Saxony, recent informative literature on these uranium ores has been almost non-existent and this gap still remains unfilled. Although much of the book is concerned with cycles of mineralization and regeneration of ores, some of the latest geochronological work is not mentioned and in 160 references (many irrelevant), several of the more outstanding German studies on the Erzgebirge fail to find a place.

**Vtorichnaya Zonal'nost' Zolotorudnykh Mestorozhdenii Urala.** [Secondary zonation in the gold deposits of the Urals.] By M. N. Al'bov. 215 pp. Gosgeoltekhizdat, Moscow, 1960. Price 10r. 50k.

It has long beeen known that the auriferous deposits of disseminated and massive sulfides in the Urals display a supergene zone in which there is a marked secondary enrichment of gold. In the first part of this book Russian and Western literature on such near-surface enrichment is reviewed, with attention given to the geochemical processes involved in the migration of gold; and in the second section it is demonstrated that similar supergene concentration of gold occurs in quartz-sulfide vein deposits, usually at depths of 20–150 meters. Below the supergene zone the grade falls off rapidly. New data include comparisons of Russian lode and placer gold demonstrating the usual increase in fineness of the latter.

**Geologiya Mestorozhdenii Redkikh Elementov.** [Geology of deposits of the rare elements.] No. 9. **Novye dannye po mineralogii redkikh elementov.** [New data on the mineralogy of deposits of the rare elements.] By A. I. Ginzburg and others. Pp. 192. Price 65k. No. 10. **Titano-tantalo-nobaty, 1.** [Titano-tantalo-niobates, part I.] By A. I. Ginzburg and others. Pp. 167. Price 63k. Gosgeoltekhizdat, Moscow, 1961.

The first of these two books by some eighteen authors at the All-Union Institute of Applied Mineralogy consists of short mineralogical papers on, *inter alia*, fergusonite, stannotantalite, lithophosphate (lithium phosphate), djalmaite = uraniferous microlite, nigerite, gaseous inclusions in various minerals of spodumene pegmatites, hydrothermal beryllium minerals, ramsayite, britholite and alumobritholite, gadolinite, rare-earth miserite and rare-earth pyrochlore, with a fund of new data. Vol. 10 is the most complete account yet published of the mineralogy of the many titano-tantalo-niobate minerals, but without geological information which presumably is to form the subject of a later volume. Some of the species described are relatively new discoveries, largely from alkalic rocks and carbonatites, and useful bibliographies of recent Russian and other papers are included.

C. F. DAVIDSON

UNIVERSITY OF ST. ANDREWS,
SCOTLAND.
*15 September, 1961*

**Structural Geology of Canadian Ore Deposits, Vol. II, Congress Volume.** Canadian Institute of Mining and Metallurgy, Montreal, Canada.

The first volume under this name was published in 1948 as a Jubilee Volume prepared by the Geology Division of the C.I.M.M. Its great success prompted the preparation of this second companion volume, also by the Geology Division. The objectives of Volume II were (1) to cover the mines or districts found or developed since 1948; (2) to describe some omitted in Volume I; and (3) to revise some of the earlier descriptions.

This volume consists of 62 papers by chosen authors, of which the largest group relate to the developments since 1948. The second group brings up to date some mines with important current production, and the third group adds, by new authors, information regarding mines described in Volume I.

As before, the deposits are treated under the Cordilleran region, the Precambrian Shield (Western, Ontario, Quebec), and Appalachian region. Each of these are divided into suitable areas. Many new mining developments have taken place in Canada since 1948, and this volume therefore contains much new information regarding Canadian ore deposits. Noteworthy are the extensive newer developments in the uranium areas of Ontario and Saskatchewan, the iron-ore deposits of Newfoundland-Quebec, the Sudbury area, Pine Point, the Chibougamou area, Lynn Lake nickel areas, and the New Brunswick deposits.

Like its predecessor volume, this Volume II will prove to be a valuable reference to many on Canadian ore deposits.

**Lehrbuch der Angewandten Geologie** (Textbook of Applied Geology). Edited by A. BENTZ. First volume: General Methods; mapping, petrography, paleontology, geophysics, soil science. Pp. 1071; figs. 468; tables 75; plates 3. Ferdinand Enke Verlag Stuttgart. 1961. Price, paperbound DM. 139.–, clothbound DM. 145.–.

This work is a veritable encyclopedia of the basic methods of geological and several related sciences. Short, numbered paragraphs, composed by dedicated specialists, deal with the individual subjects. The first part covers geomorphology and geological mapping, and here the reader gets accurate and concise information on, among other things, map-symbols, field and boring instruments, photogeology, oceanographic sampling apparatuses and mapping methods. Characteristic plants,

revealing information about the subsoil, the use of composition, size- and orientation diagrams, microtectonical methods, as well as well-arranged comparative terminological tables are also included.

The first two chapters of the book describe the modern ways of investigating different mineral and rock types and the construction of geological maps. Microchemical reactions are included in the section of geochemistry, but isotope-geology is relegated to the geophysical chapter. Very little space was assigned to structural geology and tectonical problems. Two paleontological chapters give the most important facts about the principles and methods of the up-to-date preparation, conservation and evaluation of fossil plants and animals, a most instructive and interesting reading, especially for all the "outsiders" of these fields. Although the practical aspects are stressed in this case too, enough theoretical explanation is given for its proper understanding.

Fifty percent of the work is occupied by the geophysical section. Its main features: gravimetry, geomagnetism, seismical, geoelectric and radioactive research, geothermal methods, electric- and radioactive logging, and soil-porosity measurements. The last chapter is a very short summary of the theoretical and descriptive part of soil science. In the book the sequence of topics is not illogical; it is based upon the succession of the usual phases of geologic research work. The second chapter needs a much longer list of references. As a whole the illustrative material is appropriate.

In this age of overwhelmingly increasing numbers of new scientific achievements and publications the serious threat of professional isolation often begins already in the colleges. Such outstanding, condensed, consummate works perform a vital role in fighting this trend.

ERVIN G. OTVOS

YALE UNIVERSITY,
NEW HAVEN, CONN.,
*Sept. 29, 1961*

## BOOKS RECEIVED

ROGER L. AMES AND ERIC S. CHENEY

**A Glossary of Geographical Terms.** Edited by L. DUDLEY STAMP. Pp. 539. John Wiley & Sons, New York, 1961. Price, $10.00. *An extensive glossary with appendices on Greek and Latin roots commonly used and a list of words in foreign languages absorbed into English.*

**The Long-Range Demand for Scientific and Technical Personnel.** Pp. 70. The National Science Foundation, Washington, D. C., publication NSF 61–65. *A growth rate of about 75 per cent over the 11 year period, 1959–1970, is indicated for all scientists as a group with the most rapid increases expected for mathematicians, physicists, metallurgists, and medical scientists.*

**Conference on Minerals Held in Ankara, Turkey, December, 1959.** Pp. 69. Central Treaty Organization, Yeni Meclis, Ankara, Turkey, 1961. *Representatives from Iran, Pakistan, and Turkey provided reports on the mineral industries in their respective countries.*

**Progress Report of the Office of the United States Economic Coordinator for CENTO Affairs—March 1959 to November 1960.** Pp. 38. *This report discusses a multitude of programs. Of interest to geologists is the short note on a chrome ore symposium held in Ankara.*

**Records of the Geological Survey 1957/1958.** Pp. 35. Price, 5 shillings. Lobatsi, Bechuanaland Protectorate, South Africa, 1961. *Four articles on hydrogeology, coal exploration, and the Precambrian Shushong Series.*

**Annual Report for the Year Ending December 31, 1960.** Pp. 267; tbls. 15. British Columbia Department of Mines and Petroleum Resources, Victoria, 1961. *A record of the progress of the mineral industry with details about individual operations, including those undertaken in the search for, exploration of, and development of mineral deposits, as well as the actual exploitation of mineral deposits.*

**Supplement to Minerals of California for 1955 through 1957.** JOSEPH MURDOCH and R. W. WEBB. Pp. 64. Price, $1.00. Supplement to Bulletin 173, California Division of Mines, San Francisco, 1960. *Additions to occurrences of minerals in California as reported in the literature and by personal communication from January 1, 1955 through December 31, 1957.*

**A Bibliography of Cyprus Geology.** L. M. BEAR. Pp. 15. Geological Survey of Cyprus, Nicosia, 1961.

**The Geology of the "Zuarungu" 1/2° Field Sheet.** R. J. MURRAY. Pp. 102; pls. 31; figs. 10; tbls. 27; maps 8, scale 1/250,000. Price, 16s. Geological Survey of Ghana Bull. 25, Accra, 1960. *A discussion of the petrogenesis of the granitic rocks with a brief description of several small gold prospects.*

**Land Use and Mineral Deposits in Hong Kong, Southern China and South-East Asia.** Pp. 32. University of Hong Kong, 1961. *A collection of abstracts of papers presented at the Golden Jubilee Congress.*

**The Geology of the Mapanza Mission Area.** R. TAVENER-SMITH. Pp. 25; pls. 6; tbls. 6; maps 2, scale 1/100,000. Price, 15s. Northern Rhodesia Geological Survey Rept. 10, Lusaka, 1961. *This report is devoted to the description of folded metasediments on the northern side of the Choma-Kaloma batholith and their relation to that feature. No economic mineral deposits have been found.*

**Geology of Townships 143 and 144.** J. A. ROBERTSON. Pp. 63; figs. 3; maps 3, scale 1"/¼ mile. Ontario Department of Mines Geological Rept. 4, Toronto, 1961. *A discussion of the geology and mineralization within the Blind River uranium field presenting evidence for a placer origin for the uranium.*

**Serviços Geologicos de Portugal Intrusões Kimberliticas da Lunda.** Pp. 115; pls. 32; maps. Memoir 5, Lisboa, 1959. *Stratigraphy of Karroo in Angola; diamond deposits.*

**Tysonite from Mutue Fides.** J. G. D. STEYN. Pp. 31; pls. 4; figs. 3. Price 35 cents. Geological Survey of South Africa Bull. 35, Pretoria, 1961. *Tysonite associated with fluorite, quartz, and probably hellandite is found in the tin deposits of Mutue Fides.*

**Geoloski Glasnik Broj 5.** Pp. 269. Institut des Recherches Géologiques, Sarajevo, 1961. *A collection of ten articles on paleontology, petrology and economic geology. The economic geology articles include coal, sedimentary barite, lead-zinc, and arsenic sulfide ore deposits in Bosnia.*

**Australian Bureau of Mineral Resources, Geology and Geophysics—Canberra, 1960–1961.**

**Rept. 53. Geology of the Howick Area, Singleton-Muswellbrook District, New South Wales.** J. J. VEEVERS. Pp. 22; pls. 4; figs. 12. *Bore holes in the Howick area penetrated Permian rocks comprising three cycles of sedimentation associated with the Tamago Coal Measures.*

**Bull. 50. Ordovician, Silurian, and Devonian Bryozoa of Australia.** J. P. Ross. Pp. 169; pls. 28; figs. 13. *This report includes sections on the stratigraphic distribution of bryozoan localities and the systematic paleontology.*

**Geological Survey of Canada—Ottawa, 1961.**

**Aeromagnetic maps,** scale 1/63,360, **Ontario,** Nos. 1110–1119. **Index maps, Man.-Ont.** 52–54; **Man.-Sask.** 52–64; **Sask.-Alberta** 72–74; **Alberta 84. Geologic maps,** scale 1/253,440, Nos. 4, 11 (**B.C.**), 9 (**Man.**), 10, 16 (**Sask.**), 17 (**Newfoundland**), 21 (**Man.**), 26 (**Ont.**), 34 (**P.E.I.**). **Geologic map,** scale 1/53,360, **Collins Point,** Man.

**Bull. 76. Spores with Proximal Radial Pattern from the Devonian of Canada.** D. C. McGregor. Pp. 10; pl. 1; tbl. 1. Price, 75 cents. *This report describes the morphology and occurrence of a group of spores of Devonian age.*

**Geological Survey of Fiji—Suva, 1960–1961.**

**Bull. 7. Geology of the Nandi Area Western Viti Levu.** R. W. Bartholomew. Pp. 20; pls. 14; map, scale 1/50,000. Price, 12s 6d. *Upper Eocene deposits were discovered during this study which permit correlation with similar deposits in the Marianas and Marshall islands.*

**Bull. 9. The Geology of Ovalau, Moturiki and Naingani.** P. Ibbotson. Pp. 7; map 1, scale 1/50,000. Price, 5s. *The study of this andesitic volcanic island has revealed no valuable minerals or metals.*

**Long Report no. 69. Major Extensions of the Colo-I-Suva Orebody, Determined From Self-Potentials.** R. E. Houtz. Pp. 5. *The present study suggests the orebody is essentially a flat sheet associated with basalt sills, capped by rhizolite.*

**Illinois Geological Survey—Urbana, 1961.**

**Circ. 315. Electrokinetics. III. Surface Conductance and the Conductive Solids Effect.** Norman Street. Pp. 16, appendices, 2. *This discussion concerns the theory and measurement of surface conductance, the condition that is mainly responsible for the conductive solids effect. The application of the theory to logging problems in the petroleum industry is indicated.*

**Circ. 316. Geology and Petrology of the Trivoli Sandstone in the Illinois Basin.** M. J Andresen. Pp. 31; pl. 1; figs. 15. *The Trivoli sandstone, like other Pennsylvanian sandstones, represents valley filling in an older terrane, and their basal surfaces are erosional contacts rather than facies changes.*

**Indiana Geological Survey—Bloomington, 1960.**

**Bull. 19. Fenestrate Bryozoans from the Glen Dean Limestone (Middle Chester) of Southern Indiana.** John Utgaard and T. G. Perry. Pp. 32; pls. 6; figs. 12. Price, $1.25. *Taxonomic descriptions of the fenestrate bryozoans.*

**Bull. 20. Geology of the Huron Area, South-Central Indiana.** H. H. Gray, R. D. Jenkins, R. M. Weidman. Pp. 73; pls. 3; figs. 4; tbls. 7; map, 1, scale 1/48,000. Price, $2.50. *Nonmetallic deposits in the area have been exploited in the past, but gypsum is the only product currently being worked.*

**Nyasaland Protectorate Geological Survey—Zomba, 1960–1961.**

**Bull. 12. The Geology of the Lake Chilwa Area.** M. S. Garson. Pp. 65; pls. 3; figs. 4, maps 2, scales 1/100,000 and 1/10,000. Price, 10/6. *Minerals of economic interest in the Lake Chilwa area include pyrochlore, apatite, manganese and iron-rich ores, radioactive minerals, fluorite and sulfides of iron and copper.*

**Records of the Geological Survey of Nyasaland Vol. I. 1959.** Pp. 100; pls. 4; Price, 15s–0d. *This volume is the first in a series of annual publications covering the geologic activity within Nyasaland. The reports include regional mapping, mineralogy and petrology, and economic geology.*

**Annual Report of the Geological Survey Department for the Year Ended 31st December 1960.** Pp. 19. Price, 3s–0d.

**Pennsylvania Geological Survey—Harrisburg, 1961.**

**Progress Rept. 156. Subsurface Cambro-Ordovician Stratigraphy of Northwestern Pennsylvania and Bordering States.** W. R. WAGNER. Pp. 19; pl. 1; fig. 1. *An attempt is made to reconcile Cambro-Ordovician nomenclatural sequences converging on northwestern Pennsylvania from Ohio, New York, and central Pennsylvania.*

**Progress Rept. 158. Oil and Gas Developments in Pennsylvania in 1960.** W. S. LYTLE, J. M. BERGSTEN, A. S. CATE, and W. R. WAGNER. Pp. 31; pls. 7; figs. 5; tbls. 12. *Exploration in Pennsylvania during 1960 resulted in the discovery of one new gas field, 8 new gas pools, and one deeper gas pool.*

**Information Circ. 41. Geologic Interpretation of Certain Aeromagnetic Maps of Lancaster, Berks. & Lebanon Counties.** A. A. SOCOLOW. Pp. 18. *Several magnetic anomalies in the Cornwall area are discussed in terms of magnetite ore bodies long known and mined.*

**Information Circ. 43. The Mineral Industry of Pennsylvania in 1959.** R. D. THOMSON, M. E. OTTE, and R. E. ELA. Pp. 48; tbls. 14. *Production of minerals in Pennsylvania in 1959 continued to decline; total value dropped $18.2 million, or 2 percent below 1958.*

**Atlas 127. Geologic Atlas of Pennsylvania. Geology and Mineral Resources of the Loysville Quadrangle.** J. T. MILLER. Pp. 45; pls. 2; figs. 8. Map, scale 1/24,000. *A preliminary report of the stratigraphy and structure of the Loysville quadrangle within the southern part of the Ridge and Valley province of central Pennsylvania.*

**Polish Academy of Sciences—Krakow, 1961.**

**Transactions 2. The Application of the Notion Residual Gravity to the Method of Isolating Local Anomalies from the Observed Gravity Field and Its Comparing with Certain Existing Methods.** ZBIGNIEW FAJKLEWICZ. Pp. 121. *A twelve page English summary describes the analytical methods and practical application of gravimetric surveying. In Polish.*

**Transactions 3. Copper Mineralization of the Carpathian Flysch.** H. GRUSZCZYK and B. OSTROWICKI. *In a five page English summary the authors conclude that the copper minerals had a sedimentary origin and were concentrated into lenticular bodies by a secondary process.* **On the Origin of the Copper Ore Deposit at Miedziana Góra Near Kielce.** K. PIEKARSKI. *A sedimentary origin of the Miedziana Gora mineralization is supported by the mineral association with clayey deposits, bedded character of the deposit, oxidation zone typical of other sedimentary copper deposits, and the presence of ore sulfides in unweathered gray, bituminous clay shales.*

**South Dakota Geological Survey—Vermillion, 1961.**

**Rept. of Investigations 90. Shallow Ground Water Resources in the Wagner Area, Charles Mix and Douglas Counties, South Dakota.** I. R. WALKER. Pp. 45; pls. 2; figs. 12; tbls. 5. *The Pleistocene deposits in this area include three*

*glacial outwash bodies; the lowest of these supplies most of the presently used groundwater.*

**Misc. Investigations 3. New Miocene Formation in South Dakota.** J. C. HARKSEN, J. R. MACDONALD, and W. D. SEVON. Pp. 11; figs. 3. *A volcanic ash bed is defined between the Oligocene Brule Formation and the Miocene Monroe Creek Formation in Shannon Co., South Dakota.*

**Tanganyika Geological Survey—Dodoma, 1960–1961.**

**Bull. 32. The Crystalline Rocks of the Igawa Area.** P. E. BROWN. Pp. 23; pls. 15; tbls. 5. *A comprehensive report on the petrology and structure within the Ubendian orogenic belt, with a discussion of the granite problem.*

**Annual Report of the Geological Survey Division for 1960.** Pp. 22. Price, 3/– sh.

**Geological maps,** scale 1/125,000; **sheets Moru, Kidugallo,** Nos. 37, 184.

**Geological Survey of Uganda—Entebbe, 1960–1961.**

**Rept. 3. Explanation of the Geology of Sheet 69 (Lake Wamala).** R. J. JOHNSON. Pp. 34; pls. 3; figs. 4; tbls. 2; map 1, scale 1/100,000. Price, 15s. *A series of Precambrian granitic rocks, metasediments, and sediments are described in the Lake Wamala area.*

**Rept. 4. Explanation of the Geology of Sheet 76 (Buhwezu).** A. W. REECE. Pp. 69; pls. 7; figs. 10; tbls. 3; map 1, scale 1/100,000. Price, 20s. *A detailed description of the petrology and structure of the Precambrian Igara and Buhwezu Groups.*

**Geologic and Bouger gravity map,** scale 1/1,250,000.

**U. S. Geological Survey—Washington, D. C., 1961.**

**Bull. 1032-F. Petrography and Origin of Xenotime and Monazite Concentrations Central City District, Colorado.** E. J. YOUNG and P. K. SIMS. Pp. 26; figs. 8; tbls. 7. Price, 20 cents. *The rare-earth minerals are in the lower part of a thick layer of migmatized biotite gneiss in a sequence of rocks that have been metamorphosed to the almandine amphibolite facies.*

**Bull. 1058-H. Geology of Part of the Craig C-2 Quadrangle and Adjoining Areas, Prince of Wales Island Southeastern Alaska.** C. L. SAINSBURY. Pp. 62; pls. 7; fig. 1. Map 1, scale 1/63,360. *Deposits of copper have produced ore worth $6 million. Substantial tonnages of high grade magnetite are associated with the copper but have not yet been developed.*

**Bull. 1070-C. Distribution of Uranium in Rocks and Minerals of Mesozoic Batholiths in Western United States.** E. S. LARSEN, JR. and D. GOTTFRIED. Pp. 40; figs. 4; tbls. 38. *Uranium determination of igneous rocks shows that in most rocks the major rock-forming minerals contain the bulk of the uranium present in the rock. In general, there is an increase in the uranium content of a given mineral from the mafic to siliceous rocks.*

**Bull. 1084-K. Beryllium Content of American Coals.** T. STADNICHENKO, P. ZUBOVIC, and N. B. SHEFFEY. Pp. 40; pls. 2; figs. 5; tbls. 8. Price, 70 cents. *This study suggests that the accumulation of beryllium was a syngenetic process and the content is dependent upon source rocks and position of sample locale in original sedimentary basin.*

**Bull. 1087-I. Relation of Uranium Deposits to Tectonic Pattern of the Central Cordilleran Foreland.** F. W. OSTERWALD and B. G. DEAN. Pp. 42; pls. 7;

fig. 1. *Uranium deposits within 15 tectonic subunits of the Central Cordilleran Foreland are related to the following structural environments: 1) crests of anticlines, 2) troughs of basins, 3) flanks of uplifts where smaller structures are arranged en enchelon or parallel to the major structure, and 4) at intersections of major structures.*

**Bull. 1116-E. Index to Geophysical Abstracts 180–183, 1960.** J. W. CLARKE, D. B. VITALIANO, V. S. NEUSCHEL. Pp. 58. Price, 25 cents.

**Bull. 1121-C. Eolian Deposits of the Matanuska Valley Agricultural Area Alaska.** F. W. TRAINER. Pp. 35; figs. 6. *Continuous deposition of eolian deposits in the Matanuska Valley suggests that the major source glaciers did not melt completely after the last glaciation.*

**Prof. Paper 352-C. Effect of Sediment Characteristics on Erosion and Deposition in Ephemeral-Stream Channels.** S. A. SCHUMM. Pp. 40; figs. 24; tbls. 8. *A study of erosion and sedimentation in a semi-arid environment.*

**Prof. Paper 386-A. Temperature Rise Within Radioactive Liquid Wastes Injected Into Deep Formations.** H. E. SKIBITZKE. Pp. 8; figs. 5; tbl. 1. Price, 15 cents. *As an approximation the temperature rise within an aquifer should be equal to the total heat generated in a unit volume of aquifer divided by the product of the specific gravity of the aquifer and contained waste and its heat capacity.*

**Prof. Paper 387-A. Botanical Evidence of the Modern History of Nisqually Glacier, Washington.** R. S. SIGAFOOS and E. L. HENDRICKS. Pp. 20; figs. 15. *A description of botanical methods used to determine dates of recession of three glaciers at Mount Ranier, Washington.*

**Prof. Paper 402-A. Mechanism of Gravity Drainage and Its Relation to Specific Yield of Uniform Sands.** Pp. 12; figs. 9; tbls. 3. Price, 20 cents. *A discussion of the infiltration and drainage in uniform sands.*

**Prof. Paper 411-A. A Solution of the Differential Equation of Longitudinal Dispersion in Porous Media.** A. OGATA and R. B. BANKS. Pp. 7; figs. 3. Price, 15 cents. *A contribution to the study of fluid movement in earth materials.*

**Water-Supply Paper 1475-D. Geology and Occurrence of Ground Water at Jewel Cave, National Monument, South Dakota.** C. F. DYER. Pp. 19; figs. 2. Price, 15 cents. *Bedrock in the area ranges from Precambrian to Pennsylvanian with groundwater available only from a thin sandstone bed beneath the Mississippian Englewood Formation.*

**Water-Supply Paper 1490. Geology and Ground-Water Resources of Platte County, Wyoming.** D. A. MORRIS, H. M. BABCOCK, and R. H. LANGFORD. Pp. 191; pls. 4; figs. 20; tbls. 11. *Tertiary and Recent rocks are exposed over most of the region. The aquifers of greatest potential for ground-water development are the flood-plain deposits along the North Platte River, the Arikaree Formation, and terrace deposits.*

**Water-Supply Paper 1498-D. Effect of Depth of Flow on Discharge of Bed Material.** B. R. COLBY. Pp. -2; figs. 4. Price, 15 cents. *Theoretical and empirical data show a conplex effect of depth of flow on the relationship of discharge of bed material to mean velocity.*

**Water-Supply Paper 1499-A. Water Resources of the Providence Area Rhode Island.** H. N. HALBERG, C. E. KNOX, and F. H. PAUSZEK. Pp. 49; pl. 1; figs. 16; tbls. 8. *Ground water supplies are available from glacial outwash deposits but additional smaller supplies are found in glacial till and the bedrock.*

**Water-Supply Paper 1519. Geology and Ground-Water Resources of the Owl Creek Area, Hot Springs County, Wyoming.** D. W. BERRY and R. T. LITTLETON. Pp. 57; pls. 2; figs. 9; tbls. 6. *The exposed rocks range in age from Pennsylvanian to Recent. Surficial deposits yield limited amounts of water but the Madison and Tensleep formations bear large quantities of water.*

**Water-Supply Paper 1536-B. Water Levels Near a Well Discharging from an Unconfined Aquifer.** I. REMSON, S. S. MCNEARY, and J. R. RANDOLPH. Pp. 13; figs. 2; tbls. 2. Price, 15 cents. *A numerical solution of the differential equation for unsteady flow in an unconfined aquifer.*

**Water-Supply Paper 1536-C. A simple Method for Determining Specific Yield from Pumping Tests.** Pp. 7; fig. 1. Price, 15 cents. *The new method consists of computing the volume of dewatered material in the cone of depression and comparing it with the total volume of discharged water.*

**Water-Supply Paper 1539-C. Ground-Water Reconnaissance of Winnemucca Lake, Valley Pershing, and Washoe Counties Nevada.** C. P. ZONES. Pp. 18; pl. 1; fig. 1; tbl. 1. Price 45 cents. *Ground water of good chemical quality is found only under the piedmont slopes and appears to be insufficient to sustain prolonged large withdrawals.*

**Water-Supply Paper 1539-F. Aquifers in Melt-Water Channels Along the Southwest Flank of the Des Moines Lobe, Lyon County, Minnesota.** R. SCHNEIDER and H. G. RODIS. Pp. 11; figs. 8. Price 15 cents. *By studying geomorphic limestones analyzing water-well data and drilling test holes, several elongate bodies of buried outwash having little or no topographic expression were located.*

**Water-Supply Paper 1581. Ground-Water Potentialities in the Crescent Valley, Eureka and Lander Counties, Nevada.** C. P. ZONES. Pp. 50; pl. 1; figs. 6; tbls. 5. *Fans on the west side of the valley appear capable of supplying sufficient water for irrigation with no consequent lowering of the water table.*

**Water-Supply Paper 1586-A. Water Quality and Hydrology in the Fort Belvoir Area, Virginia,** 1954–55. C. N. DURFOR. Pp. 55; figs. 20; tbls. 6. Price, 25 cents. *The quality of water in the Potomac in the Belvoir area is affected by variation in streamflow, tidal effects, and pollution.*

**Water-Supply Paper 1592-A. Resistance Coefficients and Velocity Distribution, Smooth Rectangular Channel.** H. J. TRACY and C. M. LESTER. Pp. 18; figs. 11; tbls. 2. *Laboratory studies of open-channel flow.*

**Water-Supply Paper 1610-A. Waterpower Resources of the Bradley River Basin, Kenai Peninsula, Alaska.** F. A. JOHNSON. Pp. 25; pl. 1; figs. 4; tbls. 9. *Potential power of Bradley Lake and related geographic, climatic, and hydrologic factors.*

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# SOCIETY OF ECONOMIC GEOLOGISTS

## FEBRUARY, 1962 JOINT MEETING WITH AIME

The annual meeting of the Society of Economic Geologists will be held in New York City, New York, February 19–22, 1962, in conjunction with the American Institute of Mining, Metallurgical and Petroleum Engineers. Headquarters for the meeting will be the *Statler Hotel.*

The Society luncheon is scheduled for 12:30 P.M. on Thursday, February 22, in the Ivy Suite, Mezzanine floor. Bar facilities will be available prior to the luncheon. Luncheon tickets at $4.50 each, may be purchased from the AIME in advance or at the Convention Registration Desk.

The Presidential address will be given by Philip J. Shenon at 11:00 A.M., Thursday, February 22, in the Ivy Suite. The subject will be *Economic Geology Today and Tomorrow.*

The Society will hold two general technical sessions. In addition, the Society and the Mining and Exploration Division of AIME will sponsor two symposia that should be of special interest to the members, one on *Epigenesis vs. Syngenesis,* the other on *Major and Minor Elements in the Host Rocks of Ore Bodies.*

Details of the program schedules and abstracts of the papers to be presented at the two general technical sessions are given below.

EUGENE N. CAMERON
Secretary

General Geology: Monday, February 19, 9:30–12:00 A.M., Penn Top North Room

Chairmen: to be announced.

1. W. R. Griffitts: Prospecting for beryllium.
2. Lorraine E. Patten: Field determination of beryllium contents of soils and rocks.
3. Wm. C. Kelly and E. N. Goddard: Mineralogy and paragenesis of the Telluride ores of Boulder County, Colorado.
4. P. L. Cloke and Wm. C. Kelly: An experimental study of the supergene transport of gold.
5. John S. Brown: Ore leads and isotopes.

General Geology: Monday, February 19, 2:30–4:00 P.M., Penn Top South Room

Co-chairmen: to be announced.

1. B. L. Berman: Sedimentary facies in relation to the Clinton iron ores of eastern United States.
2. Eugene N. Cameron: Origin of certain magnetite-bearing bodies in the eastern part of the Bushveld Complex.
3. Jack V. Everett and Henry Lepp: Iron deposits of the Quartz Lake area, Mt. Wright region, Quebec.

*Epigenesis vs. Syngenesis:* Wednesday, February 21, 9:00–12:00 A.M., Georgian Room

Co-chairmen: to be announced.

1. J. J. Brummer: A discussion of some syngenetic sulphide deposits.
2. T. S. Lovering: Epigenetic and diplogenetic deposits in sediments.
3. Charles Meyer: An appraisal of the evidence from "conformable" ores and sediments.
4. Discussion by Reno H. Sales, John D. Ridge, and M. L. Jensen.

*Major and Minor Elements in the Host Rocks of Ore Bodies:* Thursday, February 22, 9:00 A.M., Georgian Room

Co-chairmen: R. J. P. Lyon and H. Bloom

1. Donald B. Tallock: Redistribution of K, Na, and Al in some felsic rocks in Nevada and Sweden.
2. A. E. J. Engel and Celeste G. Engel: Adirondack orogenesis and mineral deposits.
3. A. T. Miesch: Compositions of sandstone host rocks of uranium deposits.
4. M. A. Klugman,* D. N. Bloom, D. N. Stevens: Mg/Ca ratios in carbonate wall-rock.
5. C. Wayne Burnham: Some compositional effects of hydrothermal alteration.

## ABSTRACTS OF PAPERS—GENERAL GEOLOGY SESSIONS

### *PROSPECTING FOR BERYLLIUM*

W. R. GRIFFITTS

*U. S. Geological Survey, Denver, Colo.*

Prospecting for nonpegmatitic deposits of beryllium has been handicapped by the lack of truly diagnostic geologic features and by the inconspicuousness of most beryllium-bearing minerals. The use of analyses of rock samples taken during reconnaissance examinations of promising districts is rather costly and probably is more likely to miss than to detect the beryllium deposits because even in districts that are known to contain important beryllium deposits the percentage of the total outcrop area that contains beryllium minerals is small. In the Iron Mountain area, New Mexico, for example, only 2 to 3 percent of the tactite was mapped as mineralized and probably less than half of this tactite contains beryllium. In the Lake George area, Colorado, the percent of outcrops that contain beryllium is even smaller. The nuclear devices that have been available during the last few years permit the economic checking of many more exposures than was possible with chemical analyses. Nevertheless their use has not appreciably reduced the field time needed to evaluate a district.

Evaluation of a district by the use of some characteristic of a readily sampled material is one general objective of geochemical prospecting. The beryllium content of alluvium from short streams or washes can be used to determine the presence of beryllium-rich rocks in the drainage basins upstream from the sampled localities, regardless of the nature of the beryllium minerals present. Examples are given of beryllium contents of alluvium and soils in several beryllium districts, in which beryllium determinations were made either spectrographically or by a recently developed field chemical procedure.

* Indicates speaker.

## FIELD DETERMINATION OF BERYLLIUM CONTENTS OF SOILS AND ROCKS

LORRAINE E. PATTEN
*U. S. Geological Survey, Denver, Colo.*

In the search for beryllium minerals, detection of small to moderate amounts of beryllium in soils and rocks has become increasingly important. Rapid and sensitive determinations can be made by fusion of the sample with ammonium bifluoride, followed by extraction of the beryllium with morin to form a fluorescent complex. Chelating agents and an alkaline buffer are added to minimize interferences.

The ammonium bifluoride fusion is effective on minerals that contain only minor amounts of Be as well as on beryllium minerals. The method is useful in the range from 1 to 100 ppm of beryllium in the sample, but with proper adjustment of the sample size or volumes of solution taken for analysis, the range can be extended. As many as 60 samples can be analyzed per day by an experienced analyst.

## MINERALOGY AND PARAGENESIS OF THE TELLURIDE ORES OF BOULDER COUNTY, COLORADO

WM. C. KELLY AND E. N. GODDARD
*Department of Geology and Mineralogy, Univ. of Mich., Ann Arbor, Mich.*

The gold telluride ores of the Jamestown, Gold Hill, and Magnolia mining districts have been investigated by microscopic and X-ray methods. A variety of opaque minerals including altaite, calaverite, coloradoite, empressite, gold, hessite, krennerite, melonite, petzite, rickardite, sylvanite, tellurium, and tetradymite have been recognized in a suite of 195 polished sections. In all of the ores an intimate intergrowth of two or more tellurides is characteristic. The textures suggest the growth of some tellurides in silica gel which subsequently crystallized to form horn quartz.

Experimental work in the system Au-Ag-Te reported in 1961 by Markham provides a basis for evaluating the degree of equilibrium maintained during ore deposition. Individual veins or groups of veins behaved as separate open systems in which the number of tellurides deposited commonly exceeded the number permitted by the Phase Rule. At any point in these veins, however, equilibrium was approached to the extent that only compatible phases are found in contact. These compatible phases commonly vein and replace one another and, where the textural evidence is clear, there is a definite telluride sequence involving an enrichment in silver and depletion of tellurium.

Gold is residually enriched by supergene processes but there is no extensive transport of the metal. Rickardite ($Cu_2Te$) occurs exclusively as a supergene mineral and in one unusual occurrence petzite was observed as a supergene alteration product of native tellurium. In general, the hypogene relationships are not obscured, as at Kalgoorlie, by the occurrence of some tellurides as both hypogene and supergene minerals.

## AN EXPERIMENTAL STUDY OF SUPERGENE TRANSPORT OF GOLD

PAUL L. CLOKE AND WILLIAM C. KELLY
*Department of Geology and Mineralogy, Univ. of Mich., Ann Arbor, Mich.*

Previous experiments and calculations (Kelly and Cloke, 1961; Krauskopf, 1951) have defined the ranges of acidity, chloride ion activity, and oxidation potential within which gold should dissolve. Solution 0.5 m, in $H_2SO_4$ and 0.1 M in NaCl reached a high Eh when in contact with goethite or pyrolusite and were theoretically able to dissolve gold as the $AuCl_4^-$ ion.

In the present experiments the $H_2SO_4$–NaCl solution was allowed to percolate through a column of pulverized goethite or pyrolusite and then to flow over a weighed gold plate. The concentrations of acid (1.0 N) and NaCl (0.1 N) were chosen to correspond to the strongest likely to exist in the gossans overlying ore deposits. When the solution passed over pyrolusite, the gold plate was strongly etched and lost weight. In one week's time about 50 ml of solution flowed over the plate and dissolved 2 mg of gold. More dilute solutions, of course, would have less effect. About 200 ml of the solution was passed over goethite with no noticeable effect in one month. Nevertheless, a longer experiment might be effective.

It was concluded that the rate of solution of gold is faster when pyrolusite is used instead of goethite. This corresponds to a greater theoretical solubility in the presence of pyrolusite than with goethite and to the field observation that pyrolusite is more effective than goethite in producing supergene enrichment of gold.

## *ORE LEADS AND ISOTOPES*

JOHN S. BROWN
*942 Dulaney Valley Rd., Towson 4, Md.*

Comparison of all published analyses of commercial lead deposits, excluding Mississippi Valley ores, shows that the ratio of thorium-derived to uranium-derived lead (206 + 207) has remained constant at 1.13 to 1.14 for the last 2.5 b.y. Higher figures (1.15 +) denote contamination by thorium lead and lower values indicate admixture of uranium lead. Mississippi Valley ores (1.10 ±) are contaminated by uranium lead presumably derived for local ground water and host rock. The amount of contaminant is approximately 1.5% of total lead or 3% of the uranogenic lead.

Using this stable index one may reconstruct any simple anomalous lead and determine its original ore base. Complex contamination by both thorium and uranium lead usually can be unravelled by the aid of 206/204 and 208/204 ratios appropriate to the indicated percentage of 204. This method suggests that many anomalous leads of minor veinlets in the Canadian shield are contaminated Mississippi Valley leads, probably leakages from deposits once widespread in overlying Paleozoic strata now denuded. Similarly, anomalous leads of the Baltic area indicate a former extensive Mississippi Valley stage mineralization for Europe. This seems to be still preserved in the original at Laisvall, Sweden.

Ratios of 206 and 207 to each other, especially in post-Precambrian time, are too delicately balanced and too strongly influenced by minor variables to have any reliability in age determination, but the percentage of Pb 204 is still a very useful index.

## *SEDIMENTARY FACIES IN RELATION TO THE CLINTON IRON ORES OF THE EASTERN UNITED STATES*

BYRD L. BERMAN
*Villanova University and the Ore Genesis Laboratory, Columbia University*

Analysis of the Silurian stratigraphy of the eastern United States shows that iron concentration is present throughout the Appalachian Basin east of the Nashville-Cincinnati structural lineament, an area some 200 miles in width and 950 miles in length. Ore zones cut across sedimentary facies, indicating (1) time equivalence of such zones and (2) independence of concentration process (but not amount of concentration) to any given sedimentary facies. Development of iron concentration is largely a result of diastemic conditions; development of economic deposits is a function of the character (in large part, that of grain size) of the bottom sediments during such periods. Ore development is most pronounced in those sections containing the greatest thickness of coarse clastics, either quartz sand or

bioclastics. These concepts allow the preparation of lithofacies maps outlining those areas in which commercial quantities of ore can be expected. This is illustrated by such a map drawn for the Red Mountain Formation of Alabama.

## *ORIGIN OF CERTAIN MAGNETITE-BEARING BODIES IN THE EASTERN PART OF THE BUSHVELD COMPLEX*

EUGENE N. CAMERON
*Science Hall, University of Wisconsin, Madison 6, Wisconsin*

Bodies of pegmatitic rock consisting of clinopyroxene, plagioclase, amphibole, and magnetite, with other minerals as yet unidentified, occur at a number of places in the Critical Zone of the Bushveld Complex in the Eastern Transvaal. The bodies occupy areas ranging from a few hundred square feet to nearly a square mile. They are of particular interest because they carry vanadium-bearing magnetite and because they may be related to the well known platinum-bearing pipes of this part of the Bushveld Complex.

Studies of such bodies on Farms Annex Grootboom, Grootboom, and Tweefontein indicate that they are developed largely in the anorthosite series forming the upper part of the Critical Zone. The anorthosite series is largely a pseudostratified succession of anorthosites and noritic anorthosites, but persistent layers of pyroxenite and chromitite occur at several horizons. The chromitite seams can be traced through the pegmatitic bodies; locally the chromitites have been partially replaced by magnetite. The pyroxenite layers can likewise be traced through the pegmatitic bodies, although locally pyroxenite is likewise partly replaced. It is evident that the pegmatitic bodies are not intrusive but have formed by replacement, largely at the expense of anorthositic rocks. The development of the bodies has been accomplished without disruption of the gross layered structure of the rocks of the Critical Zone.

## *IRON DEPOSITS OF THE QUARTZ LAKE AREA, MT. WRIGHT REGION, QUEBEC*

JACK V. EVERETT AND HENRY LEPP
*400 Torrey Building, Duluth 2, Minnesota*

The essentially unmetamorphosed iron formations of the Labrador geosyncline cross the boundary of the Grenville metamorphic province approximately 15 miles north of Wabush Lake, Labrador and extend within this geologic province for about 200 miles southwesterly to Matonipi Lake, Quebec. In the Grenville the iron formations are highly metamorphosed, folded and discontinuous. The purpose of this paper is to describe two iron deposits which are currently being explored near the northeast end of the belt at Quartz Lake (also known as Moire Lake) Quebec.

The principal rock types in the area are granite gneiss, biotite and garnet schists, amphibolites, quartzite and iron formation (including quartz-magnetite-specularite, quartz-specularite, quartz-magnetite-silicate, and quartz-silicate varieties.

Interpretation of the sequence of formations and the geologic structures is complicated by the high degree of recrystallization, by the locally intensive plastic deformation, and by the fact that there is considerable interfingering of the iron formation members, and the amphibolites. The major structures in the area appear to be doubly plunging synclines.

The favorable oxide members of the iron formation contain about 30% soluble iron. Concentration tests show that this material can be concentrated to a product containing approximately 66% iron and 5% silica, with a 40% weight recovery. Generally, the specular hematite is liberated with a −35 mesh grind, the magnetite with a −100 mesh grind.

Preliminary studies indicate that substantial reserves of concentratable iron formation exist in the Quartz Lake area.

*Economic Geology*
Vol. 56, 1961, p. 1496

# SCIENTIFIC NOTES AND NEWS

GEORGES ORDONEZ has resigned from Compañia Minera Kenmex, S. A. (Kennecott Copper Corp.) and has opened a consulting office for exploration, evaluation, and development of mining properties at Avenida Juarez 134–403 Mexico 1, D. F.

The 11th Clay Minerals Conference is to be held 13–17 August, 1962 at Ottawa, Ontario, Canada.

The 1962 recipient of the Hal Williams Hardinge Award of the A.I.M.E. will be IAN CAMPBELL, California State Geologist.

E. W. WHEELER, after 5 years as geologist for Andes Copper Co., has taken a similar position with Bear Creek Mining Co. in Tombstone, Arizona.

CORRECTION: On pages 877–878 Fig. 2 should be Fig. 3, and Fig. 3 should be Fig. 2.

LLOYD E. ELKINS of Tulsa has been elected President of the American Institute of Mining, Metallurgical, and Petroleum Engineers for 1962. He will take office February 20 during the 91st Annual Meeting of the Institute, to be held in New York.

CORRECTION: On page 987 "Minargraga" should read Minasragra; "Iquitos" should be deleted. On page 990 the figures of national income should read 9.5 and 3.8 instead of "11" and "4." Acknowledgent should be made to Mr. John Burgess, Jr., and Mr. Ulrich Peterson.

LAUREN A. WRIGHT, formerly Supervising Mining Geologist with the California Division of Mines, has joined the faculty of the Pennsylvania State University as Professor of Geology and Head of the Department of Geology. He succeeds FRANK M. SWARTZ, who has been appointed Research Professor of Paleontology. PETER J. WYLLIE, returning to Penn State from Leeds University, joins the Department of Mineralogy and Petrology as Associate Professor of Petrology. RICHARD R. PARIZEK has come from the University of Illinois as Assistant Professor of Geology, and will activate a program of instruction and research in ground water geology. Scheduled to arrive early in 1962 are CECIL E. TILLEY of the University of Cambridge, who will serve as Visiting Professor of Petrology, and WILLIAM G. CHALONER of University College, London University, who will serve as Visiting Professor of Palynology.

TERAH L. SMILEY, director of The University of Arizona's Geochronology Laboratories, and Dr. JOHN W. HARSHBARGER, head of the U of A's Department of Geology, will serve as co-chairmen of the major symposium on "Water Improvement" during the 128th annual meeting of the American Association for the Advancement of Science December 26–30 in Denver.

The Third Caribbean Geological Conference will be held in Jamaica April 2 to 11, 1962, at the University College of the West Indies, Mona. Registration forms may be had from the Geological Survey Department, Kingston.

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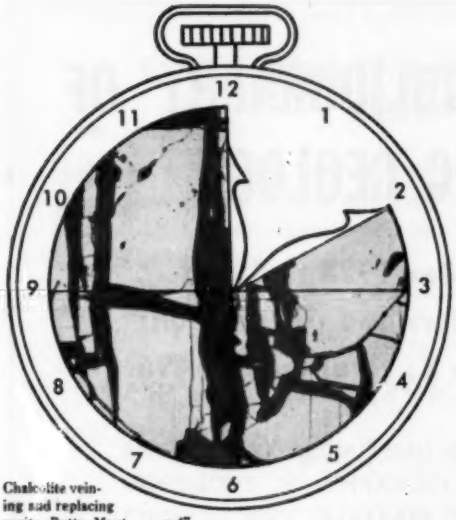


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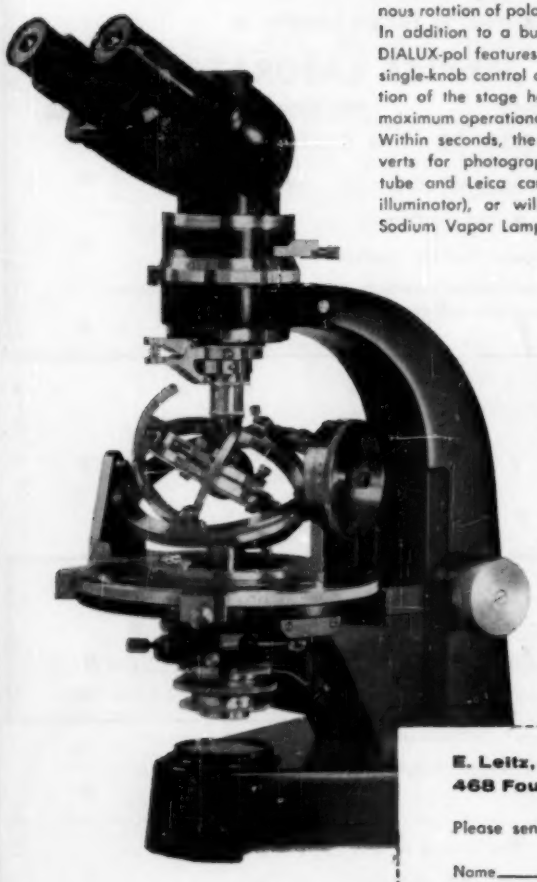

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